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Atmospheric Structure Prior to Tornadoes as Derived from Proximity and Precedent Upper Air Soundings

Covering the Period
April 1977 - June 1979

Prepared by G. E. Taylor, G. L. Darkow

Department of Atmospheric Science
University of Missouri

Prepared for
U.S. Nuclear Regulatory
Commission

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ABSTRACT

The uniqueness of the thermodynamic and dynamic structure of the atmosphere in the area of imminent tornado bearing storm development is analyzed by comparing 115 tornado proximity soundings with upper air soundings made at the same location 6 and 12 hours earlier (precedent soundings) and with soundings made simultaneously at neighboring upper air stations.

The comparisons suggest that both the proximity station and the neighboring station upstream with respect to the mean flow in the low level moist air display very similar degrees of hydrostatic and potential-convective instability by late afternoon. The principal difference is in the wind profiles at the two locations. The tornado proximity station displays significantly stronger wind speeds above 1 km with the most striking difference being in the vertical shear of the wind in the layer from 1 to 3 km above ground level.

In this layer the winds at the proximity station show an average increase of about 3 m sec^{-1} while the upstream, non-tornadic, station shows a slight decrease of wind speed with height.

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CHAPTER I

INTRODUCTION

The purpose of this thesis is to report on an attempt to detect and measure the slight differences between the structure of the atmosphere in areas capable of producing tornado bearing thunderstorms and that of neighboring areas in which only non-severe thunderstorms form. This investigation is based on a comparison of upper air soundings made in the vicinity of subsequent tornado bearing storm formation (tornado proximity soundings) and soundings made simultaneously at the closest upper air station upstream with respect to the low level moist flow (check soundings). The diurnal evolution of the atmospheric structure capable of producing tornado bearing thunderstorms was examined by comparing the tornado proximity soundings with associated soundings released at the same station as, but prior to, the tornado proximity soundings (6 and 12 hour precedent soundings).

Previous investigations have failed to completely determine the subtle differences between the environment in which tornado bearing thunderstorms develop and neighboring environments in which at most only non-severe storms develop. A serious consequence of this is our inability to accurately forecast the time and location of these severe weather occurences. In light of the catastrophic nature of violent tornadoes, any attempts to detect and measure these slight differences are necessary.

Although tornadoes are one of the atmosphere's smallest vortices they are locally the most destructive. A typical tornado is characterized

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by a cylindrically shaped rotating cloud extending from the bottom of the "parent" thunderstorm to the ground. It is typically on the ground for only a few minutes and damages an area approximately 50 to 150 meters wide along a path 2 to 10 km in length. Damage produced by the tornado is usually attributed to wind speeds in excess of 35 m/sec and the sudden pressure drop of the order of 10 to 25 percent (Schaefer et al., 1979).

Tornadoes in the U.S. are responsible for an average of 125 deaths per year and property losses valued at $75 million (Kessler, 1970).

Another important characteristic of tornadoes is their extreme variability. Small tornadoes may touch down momentarily causing only minor damage to a small area. Conversely, large intense tornadoes may stay on the ground for two or more hours producing destruction along a continuous path over 200 km in length.

It is these few large tornadoes that produce most of the tornado losses (Cressman, 1969). In fact, Pearson (1971), Director of the National Severe Storms Forecast Center, has noted that slightly over 1 percent of the reported tornadoes in the period from 1960 to 1970 produced more than 80 percent of the tornado fatalities. Some of these fatalities were a direct result of our inability to effectively warn the public (Kessler, 1970).

Because of tornadoes' significant socioeconomic impact, atmospheric scientists are continually striving to improve their understanding and prediction capabilities of these events. A major problem of severe weather research has been to determine the key differences in the thermodynamic stability and the wind flow regime between an environment in which subsequent severe storm activity occurred and an environment in which at most only non-severe thunderstorms developed. Once these key

differences have been identified, coupled with the ability to forecast when and where these important differences will occur, our prediction capabilities of severe weather will be greatly enhanced.

Synoptic scale severe storm research can be categorized into three main areas: 1) analysis of the thermodynamic structure of the pre-severe storm environment; 2) analysis of the dynamic structure of the pre-severe storm environment; 3) resolution of the triggering mechanisms required to initiate intense penetrative convection.

In attempts to determine the thermodynamic structure of the environment in which subsequent severe storms occurred, many researchers have utilized radiosonde data. In one of the earlier reports of this type, Showalter and Fulks (1943) stated that a sounding typical of the environmental air in the immediate vicinity of severe storms was characterized by moist low level air capped by a strong inversion and with marked drying aloft. In an analysis of 72 tornado proximity soundings, Fawbush and Miller (1952) noted that 70 percent of their soundings had inversions. The average depth of the moist layer was 1500 meters and the relative humidity within this layer was approximately 85 percent.

In later studies, Beebe (1958, 1963) presented strong evidence that the "typical" inversion was not present in the area of and at the time of tornado genesis. Although low level inversions may have been present earlier in the day in the area of subsequent severe weather, Beebe (1958, 1963) found that the inversion had disappeared and that moisture had been pulled up to great heights in the atmosphere prior to the severe storm activity. A tornado proximity sounding analysis by Darkow (1969) tended to support Beebe's results. Darkow found that only

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3 out of 15 tornado proximity soundings released within 50 km of and 1 hour prior to the tornado had inversions. Both Beebe (1958, 1963) and Darkow (1969) found that the mean tornado sounding was characterized by potential convective instability with moisture extending to great depths in the atmosphere.

The substantial differences in the results of Fawbush and Miller (1952) as compared to the results of Beebe (1958, 1963) and Darkow (1969) were probably due to the much less stringent requirements Fawbush and Miller used to establish their tornado proximity sounding collection. Fawbush and Miller included in their collection any sounding released within 325 km and 6 hours of the tornado occurrence. In contrast, Beebe's criteria stated that the sounding must be released within 1 hour prior to the occurrence of a tornado within 80 km of the sounding release location.

In an attempt to explain why some of their tornado proximity soundings did not contain low level inversions, Fawbush and Miller (1954) stated that tornadoes occur in different types of air masses. They classified the bulk of their tornado proximity soundings into two divisions: 1) Great Plains Type; 2) Gulf Coast Type. The major difference between these two divisions is the presence of a low level capping inversion in the Great Plains Type air mass which is nonexistent in the Gulf Coast Type air mass.

Later studies (Darkow, 1969 and Wills, 1969) did not support Fawbush and Miller's results. Although Darkow (1969) and Wills (1969) did note small differences in the pre-severe storm environments between various regions in the U.S., significant differences, such as the persistent presence of low level inversions in a particular region and

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the absence of inversions in other regions, were not observed.

The dissimilar results of the aforementioned atmospheric scientists clearly demonstrates the importance of careful data selection. Obviously, the criteria utilized for tornado proximity sounding selection can significantly influence the results obtained.

Even though the results of the previously cited investigations were dissimilar in some aspects, the researchers along with Gray (1971), Reap (1977), Charba (1978) and others concur upon certain thermodynamic conditions favorable for the occurrence of severe storms and tornadoes. These conditions are (Newton, 1963): 1) availability of abundant moisture in lower levels; 2) conditional instability to great depths in the atmosphere; 3) potential-convective instability (a decrease of equivalent potential temperature or wet bulb potential temperature with height).

Given that these three conditions are important for severe storm development, it is important that we: 1) understand the destabilization process in order to forecast when and where these conditions will be satisfied; 2) determine if it is possible to delineate regions with a high potential for subsequent severe weather development by spatial analysis of these thermodynamic conditions; 3) develop techniques to quickly and accurately determine the degree of instability of the atmosphere.

Tornado proximity soundings have also been utilized in the analysis of the destabilization process. Lustig (1973) combined tornado proximity soundings with associated precedent soundings (upper air soundings released at the same station but prior to the tornado proximity soundings) to examine this process. He found that the 12 hour destabilization occurred primarily in the lowest 4 km of the atmosphere. He

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noted that two main factors contributed heavily to the destabilization process. Insolational heating and convective mixing increased the temperature and lapse rate in the lowest 1.5 km of the atmosphere. Moisture addition by advection and/or evapotranspiration of approximately 1.4 g/kg from the surface to 3 km was also important to the destabilization process. The combination of surface heating and low level moisture addition produced a substantial increase in the potential-convective instability in the lowest 4 km of the atmosphere.

Other investigators have utilized case studies to examine the destabilization process. Sugg and Foster (1954) found that advection of warm, moist air in lower levels was strong enough to decrease the stability of the atmosphere despite warm air advection at higher levels in one particular tornado situation. Conversely, Whitney and Miller (1956) concluded that differential temperature advection alone was sufficient to produce a rapid decrease of stability at the time and in the area of tornadoes. House (1959) stated that present in a typical warm air sector in which severe weather occurs is a pattern of vertical motion, temperature advection, and low level insolational heating that will produce a marked decrease of stability in a restricted area. This area is the favored location for squall line development.

In conclusion, the literature cited above shows that insolational heating, moisture changes, horizontal advection, and vertical motion are all important components of the destabilization process. If insolational heating is reduced due to cloudiness, less intense convection would be the result. Therefore, it is clear that all such components should be considered in forecasting convective phenomena (Whitney and Miller, 1956).

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To examine the subtle differences between pre-severe storm environments and non-severe storm environments, Darkow (1968, 1969) compared tornado proximity soundings to associated check soundings. A check sounding was an upper air sounding released at the same time as the tornado proximity sounding at the nearest upper air station in an upstream direction with respect to the mean flow in the low level moist layer. Furthermore, the check sounding must have been released within the same air mass as the tornado proximity sounding.

Darkow's (1968, 1969) results show negligible differences in temperature and lapse rate between the tornado proximity soundings and the check soundings. However, there was a pronounced difference in the moisture fields. The tornado proximity soundings had larger amounts of moisture extending to greater depths in the atmosphere than the associated check soundings.

From these results, one might hypothesize that spatial analyses of instability could delineate areas of subsequent severe weather development. However, not all investigators have reached similar conclusions.

Maddox (1976) also examined upper air soundings representing pre-severe storm environments and non-severe storm environments. He noted that the range of conditions associated with tornadoes was very large. Also, he determined that the differences between the mean pre-severe storm sounding and the mean non-severe storm sounding were small. In conclusion, Maddox stated that the only necessary condition on the synoptic scale for the probability of tornadoes is that convection must occur.

Once again, differences in criteria utilized to establish their

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tornado proximity sounding collections may in part explain the differences in conclusions reached by the two investigators. Maddox's tornado proximity sounding collection criteria was less restrictive than Darkow's. Less restrictive criteria would tend to increase the range of conditions associated with the pre-storm environment. Therefore, in the final analysis, the averaging process might smooth out important, subtle differences between the mean pre-severe storm sounding and the mean non-severe storm sounding.

The two investigations cited above represent opposing viewpoints regarding the utility of spatial analyses of the thermodynamic structure of the atmosphere to severe storm forecasting. However, the continued use of stability indices indicates that many meteorologists believe that spatial analyses of the degree of instability of the atmosphere can be an aid in forecasting thunderstorms and severe storms.

Numerous stability indices have been developed over the years in an attempt to quickly and accurately ascertain the degree of instability of the atmosphere. To comment on all of the stability indices is beyond the scope of this thesis. Therefore, only a few of the more widely utilized stability indices will be mentioned here.

One of the first stability indices developed was the Showalter Index (Showalter, 1953). Its numeric value is the difference between the environmental temperature at 500 millibars (mb) and the temperature of a parcel lifted adiabatically from 850 mb to 500 mb. Negative values indicate the degree of latent instability. For example, a sample of the atmosphere which has a Showalter Index value of -5 contains a significantly greater degree of latent instability than a sample whose Showalter Index value is -3. The Showalter Index has the advantages of objectivity

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and ease of calculation from routinely reported data; however, it is very sensitive to the depth of the moist layer. If the moist layer does not reach the 850 mb level, the Showalter Index will underestimate the degree of latent instability.

The Lifted Index (Galway, 1956 and Winston, 1956) uses the average temperature and mixing ratio in the boundary layer for the initial characteristics of the parcel. After lifting the parcel adiabatically to 500 mb, the difference between the environmental temperature and the parcel's temperature is the numeric value of the Lifted Index. Equivalent to the Showalter Index, negative values indicate the degree of latent instability of the atmosphere. The Lifted Index is not overly sensitive to the depth of moisture layer as is the Showalter Index; therefore, the Lifted Index is more accurate than the Showalter Index with regard to that particular aspect. However, the Lifted Index is more difficult to calculate and requires an accurate estimate of the low level temperature and moisture content to be an effective forecast tool.

Fujita, Bradbury and Van Thullenar (1970) noted that any index computed from a specified level might misrepresent the stability of an air mass. With this in mind, they developed the "Best Lifted Index." The Best Lifted Index is computed similarly to the Lifted Index. Several Lifted Index values are calculated utilizing average temperatures and moisture contents for various 50 mb increments within the boundary layer. The lowest Lifted Index value becomes the Best Lifted Index value for that air mass. Fujita, Bradbury and Van Thullenar (1970) noted that the Best Lifted Index worked well as a forecast tool for delineating regions of severe weather if advective factors were taken into consideration.

The K Index (George, 1960) is primarily applicable for the

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prediction of air mass type thunderstorms. The equation used to compute the numeric value of the K Index is

$$K = T_{850} - T_{500} + T_{d\ 850} - (T_{700} - T_{d\ 700})$$

where T_{850} = Temperature at the 850 mb level

$T_{d\ 850}$ = Dew Point Temperature at the 850 mb level

T_{500} = Temperature at the 500 mb level

T_{700} = Temperature at the 700 mb level

$T_{d\ 700}$ = Dew Point Temperature at the 700 mb level.

The unique feature of the K Index is the incorporation of the dew point depression at 700 mb in its calculation. The dew point depression was included to account for the reduction of buoyancy of thunderstorm updrafts produced by the entrainment of dry environmental air. K Index values of 35 or greater are usually indicative of the possibility of numerous thunderstorms.

Miller (1972) has found good success with the Total Total Index (TT). The equation used to compute its numeric value is

$$TT = T_{900} + T_{d\ 900} - 2 \cdot T_{500}$$

where T_{900} = Temperature at 900 meters above the surface

$T_{d\ 900}$ = Dew Point Temperature at 900 meters above the surface

T_{500} = Temperature at the 500 mb level.

Miller (1972) stated that 92 percent of the reported tornadoes in 1964-1965 occurred within air masses characterized by TT values of 50 or greater.

More recently, Charba (1978) reported that modified versions of the Lifted Index and the Total Total Index were being utilized as predictors in the regression equations for forecasting severe weather developed by the Techniques Development Laboratory of the National

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Weather Service.

The thermodynamic structure of the atmosphere is only one of the factors important to severe storm outbreaks. Early in the history of severe storm research, it was recognized that the dynamic structure of the atmosphere also played an important role in severe storm development.

Fawbush, Miller, and Starrett (1951) noted that a wind speed maximum along a relatively narrow band between 3 and 6 km above the surface was usually present in the region of severe weather activity. Newton's (1963) conditions favorable for the occurrence of tornadoes includes bands of strong winds in the lower and upper levels, generally veering with height. In a later study, Darkow and Fowler (1971) observed substantially greater wind speeds above 1 km in the tornado proximity soundings as compared to the associated check soundings.

The investigators cited above along with Reap (1976), Charba (1978) and others concur regarding the importance of strong winds to severe storm development. However, not only is wind speed important to severe weather activity, but also the vertical wind shear.

Browning and Ludlam (1962) and Browning and Donaldson (1963), in two separate case studies recognized the importance of a strongly sheared environment to long-lived intense cumulonimbus. This claim has been supported by more recent investigations.

In an analysis of over 700 tornado proximity soundings, Wills (1969) found that the vertical wind shear from the surface to 500 mb averaged 23 m/sec. From these results, Wills (1969) concluded that given the presence of instability, wind shear in the lower half of the troposphere is the most important tool for predicting tornado occurrences.

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Darkow and Fowler (1971) have also found wind shear to be a good delineator of severe storms. In a spatial analysis of tornado proximity soundings and associated check soundings, Darkow and Fowler (1971) noted significantly greater wind speed shear in the layer from 1 to 3 km above the surface in the tornado proximity soundings as compared to the check soundings.

Results such as those cited above have led to: 1) a model of tornado genesis based on a strongly sheared environment (Gray, 1969 and 1971); 2) a Severe Weather Threat Index (SWEAT Index) which incorporates not only stability criteria but also information regarding low and mid-level wind speeds and the veering of the wind in the lower half of the troposphere (Miller and Maddox, 1975).

When the thermodynamic and dynamic structure of the atmosphere is favorable for severe weather, all that is necessary for the occurrence of severe storms is the presence of a triggering mechanism capable of initiating intense penetrative convection. There have been many such triggering mechanisms proposed.

The lifting provided by cold and warm fronts are two of the more obvious triggering mechanisms. A more subtle instability release mechanism is the lifting provided by the cold dome residual from earlier thunderstorm activity. A triggering mechanism unique to the southwest plains is the "dry line", a front-like discontinuity between the moist air moving northward from the Gulf of Mexico and the dry desert air from the southwest. Beebe and Bates (1955) developed models of jet axes which can serve as triggering mechanisms. Particular intersections of low and mid-level jets will produce low level convergence and mid-level divergence resulting in sufficient lifting in certain locations.

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These are a few of the triggering mechanisms capable of initiating intense penetrative convection. It is important to note that there have been others proposed, such as those by Tepper (1950a, 1950b) and other investigators.

Although triggering mechanisms were not explored, by the use of tornado proximity sounding analysis the author has noted important aspects of the destabilization process and key differences between the pre-severe storm environment and the non-severe storm environment. It is the intended purpose of this thesis to present these results and, hopefully, to expand our knowledge and understanding of the environmental structure capable of producing severe storms.

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CHAPTER II

DATA SOURCE AND ANALYSIS

The initial sources of information for the location and time of tornadic occurrences were the monthly summaries of "Storm Data" published by EDS-NOAA. These reports were scanned for potential proximity tornadoes for the period January 1965 through August 1976. A reported tornadic occurrence within the continental U.S. was designated as a potential proximity tornado if it appeared to satisfy the screening criteria.

For these tornadoes, the associated upper air sounding and supporting data, including WSR-57 radar films and surface observation report forms, were obtained from the National Climatic Center of EDS-NOAA. With this information, each case was subjected to a thorough examination to determine if it satisfied the tornado proximity sounding collection criteria. For a tornado to be included in the collection, it must have occurred within 80 km of an upper air station and within 105 minutes after sounding release time.

The upper air soundings, the surface observations and radar films were studied in order to eliminate from the collection all cases in which the sounding was post-frontal or post-squall line and all cases in which the sounding was precipitation affected. Furthermore, those cases in which the radar films did not show precipitation echoes of sufficient intensity to substantiate the report of a severe weather incident were deleted from the collection as questionable reports.

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The resulting tornado proximity sounding collection contains 161 0000 GMT soundings with data through at least 15 km above ground level (AGL).

For temporal comparison with the tornado proximity soundings, when available, associated 6 hour and 12 hour precedent soundings were obtained. As stated earlier, precedent soundings are soundings released from the same station as but prior to the tornado proximity soundings.

In addition, it was felt that a spatial comparison might be beneficial. Therefore, a check sounding collection was generated. A check sounding is a non-tornado proximity sounding released in the same air mass as and at the same time as the associated tornado proximity sounding. Furthermore, it is the sounding released at the upper air station nearest to the tornado proximity sounding station in the upstream direction with respect to the mean flow in the low level moist layer.

In connection with the check soundings, associated 12 hour check precedent soundings were obtained. These soundings bear the same relationship to the check soundings as the precedent soundings have to the tornado proximity soundings.

The final sample contains 115 cases with data complete through at least 12 km (AGL). Each case is comprised of a tornado proximity sounding and the associated 12 hour precedent, check, and check precedent soundings.

A second sample with data complete through at least 10 km (AGL) comprised of tornado proximity soundings and the associated 6 hour and 12 hour precedent soundings totals 24 cases. The tornado proximity soundings and associated 12 hour precedent soundings in 17 of the 24 cases are contained within the larger 115 case sample. Hence, these

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17 cases can be considered a subsample of the 115 case sample. The tornado proximity soundings and associated 12 hour precedent soundings in the remaining 7 cases are not contained within the 115 case sample because these soundings do not have data complete through 12 km (AGL). Thus, in a strict sense, the 24 case sample is not a subsample of the 115 case sample.

The averages and standard deviations of certain variables were computed for both the 17 case subsample of the 115 case sample and the 24 case sample. Since the results revealed only minor differences, it was decided to present only the results from the larger 24 case sample.

Careful examination of the data within the 115 case sample revealed that the surface conditions as reported in the radiosonde data of 14 of the tornado proximity soundings did not accurately reflect the conditions of the environmental air in the near vicinity of the tornado bearing thunderstorm. This would be the case, for example, if the upper air station experienced cool moist outflow from nearby shower activity or cool moist flow advected from a nearby lake or ocean. For these cases, the surface conditions were modified by examining surface station observations from locations closer to the parent thunderstorm than the upper air station.

In addition, 8 soundings had missing winds at various levels. For these soundings, the missing values were estimated by interpolating vertically between reported levels and/or horizontally between adjacent upper air stations. It should be noted that only one tornado proximity sounding contains interpolated wind data. Furthermore, only one sounding contains interpolated wind data below 5 km (AGL) and that sounding is a check precedent sounding. It should also be noted that

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generally the interpolation was over a relatively shallow layer (in over 62 percent of the soundings with interpolated wind data, the interpolation was over a layer 4 km or less in depth).

It was decided that including the soundings with estimated wind data in this study would not substantially alter the results since the inherent error in the reported wind data is comparable with the error introduced by the interpolation of the missing values.

In their original form, the soundings contained data at the surface and at specified pressure levels above the surface. This format has one distinct disadvantage when averaging soundings with widely varying surface pressures. The boundary layer characteristics may be lost. Therefore, in order to preserve these important characteristics, the data were interpolated to a height above ground format before averaging.

After conversion to this type of format, the average tornado proximity sounding, average precedent sounding and the average check sounding were calculated. To help resolve differences between these mean soundings, the average difference and variability between pairs of soundings (a tornado proximity sounding and a precedent sounding pair, etc.) were computed. Furthermore, the average difference in the 12 hour changes between the tornado proximity sounding station and the check sounding station were calculated in order to examine the difference in daytime destabilization between the two locations.

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CHAPTER III

RESULTS OF THE 115 CASE SAMPLE

1. Comparison of the Tornado Proximity and Check Soundings

The temperatures and dew points from the surface to 12 km (AGL) for the average proximity and average check soundings are shown in figure 3.1. The lapse rates of the two temperature profiles are very similar. In the lowest kilometer of the atmosphere, the lapse rates for both soundings are approximately 8°C per km. Above this layer, the lapse rates ranged from 6 to 7°C per km. The temperatures of the two soundings are similar at corresponding altitudes; however, the average check sounding is warmer at all levels.

The dew point temperature profiles exhibit even more similarity. This is surprising in light of the results of an earlier study. Darkow (1969), in an analysis of proximity and check sounding data, found that the proximity soundings possessed substantially greater amounts of moisture in the lower levels of the atmosphere than the associated check soundings. However, the similarity of dew point temperature profiles in figure 3.1 suggests only small differences in moisture content between the two soundings.

The average difference in temperature at each level between the two sets of soundings is shown in figure 3.2. The tick marks equidistant on either side of the average values represent plus and minus one standard deviation. Identical notation will be used in subsequent figures.

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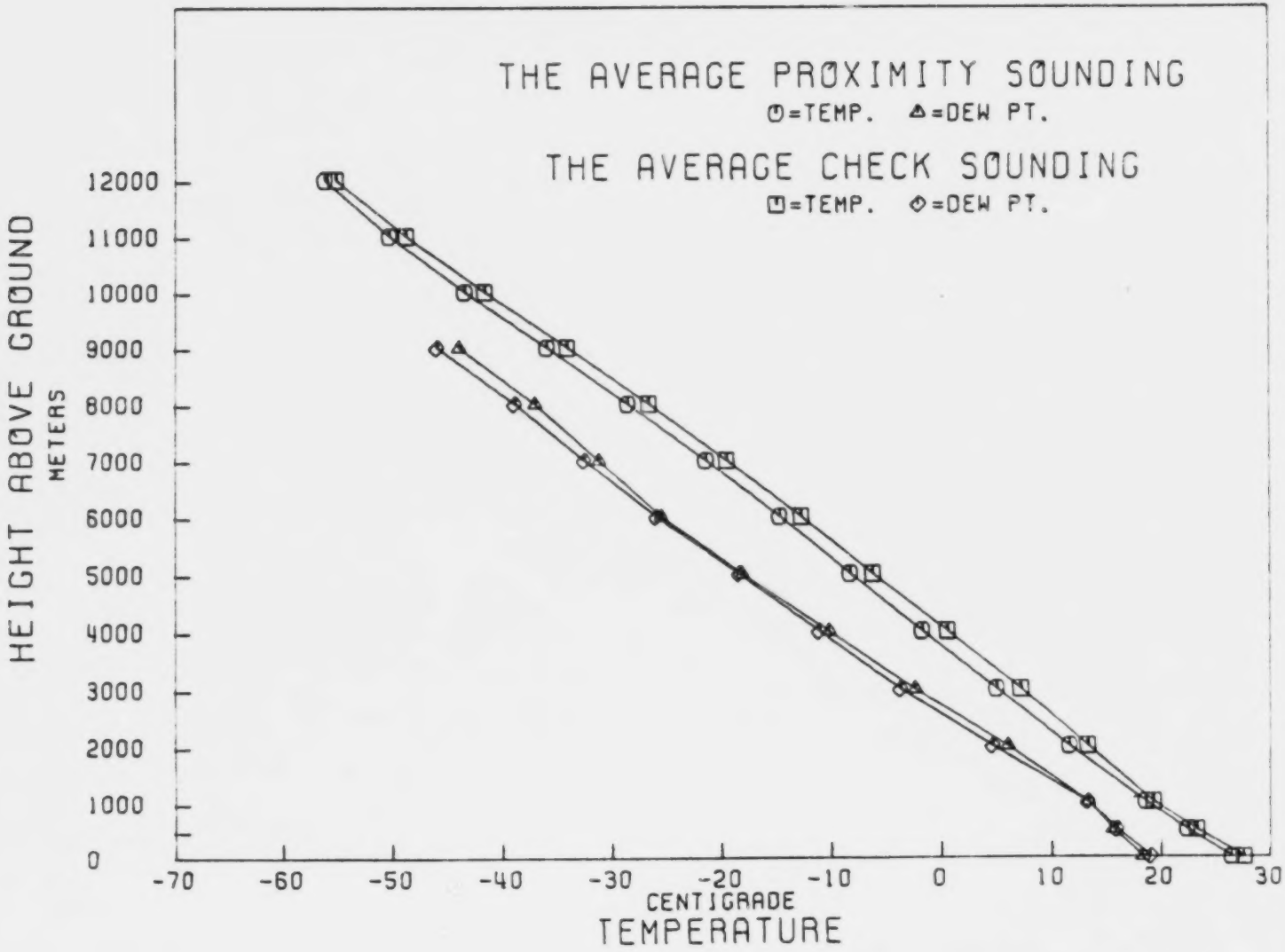

Figure 3.1 Average temperatures and dew point temperatures for the tornado proximity soundings and the associated check soundings (115 cases).

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The negative average values in figure 3.2 indicate that the average check sounding is warmer than the average proximity sounding at all levels from the surface to 12 km (AGL). From the surface to 1 km (AGL), the average check sounding is approximately $1^{o}C$ warmer than the average proximity sounding. This difference is of the same size as the generally accepted $\pm$ $1^{o}C$ accuracy in reported temperatures in rawinsonde data (Fankhauser, 1974). Above 1 km (AGL), the average check sounding is approximately $2^{o}C$ warmer except at 12 km (AGL) where the check soundings averaged $1^{o}C$ warmer than the proximity soundings. The scatter represented by the tick marks indicates considerable variability in temperature differences between the proximity and check soundings. Furthermore, the scatter indicates that above 1 km (AGL) the check soundings are warmer than the associated proximity soundings in 70 percent of the cases. At 1 km (AGL) and below, the check soundings are warmer in approximately 60 percent of the cases.

Figure 3.3 shows the average difference in mixing ratio from the surface to 9 km (AGL). The figure shows only small differences with large variabilities between the two sets of soundings. On the average, the check soundings have slightly higher mixing ratio values at the surface and at 0.5 km (AGL) than the proximity soundings. At all other levels, the mean proximity sounding possesses greater moisture amounts. The maximum average difference in moisture content occurs at 2 km (AGL). At this level, the average proximity sounding's water vapor content is 0.8 g/kg greater than the moisture content of the average check sounding. The accuracy of the humidity sensor in rawinsondes is approximately $\pm$ 10 percent for relative humidities greater than 30 percent (Fankhauser, 1974). This corresponds to an error in mixing ratio of approximately $\pm$ 1 g/kg.

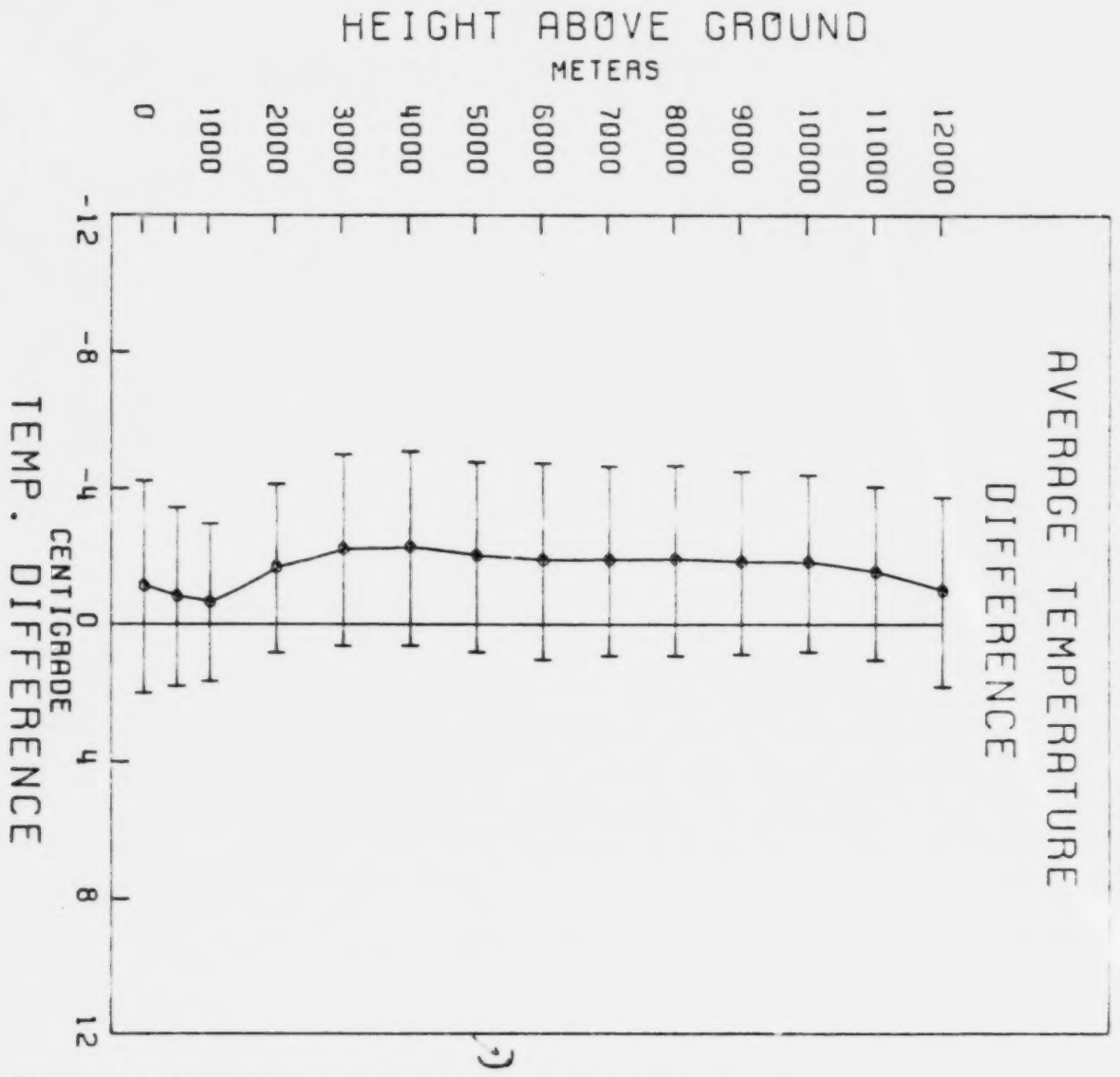

Figure 3.2 Average temperature difference between the 115 pairs of tornado proximity soundings and associated check soundings.

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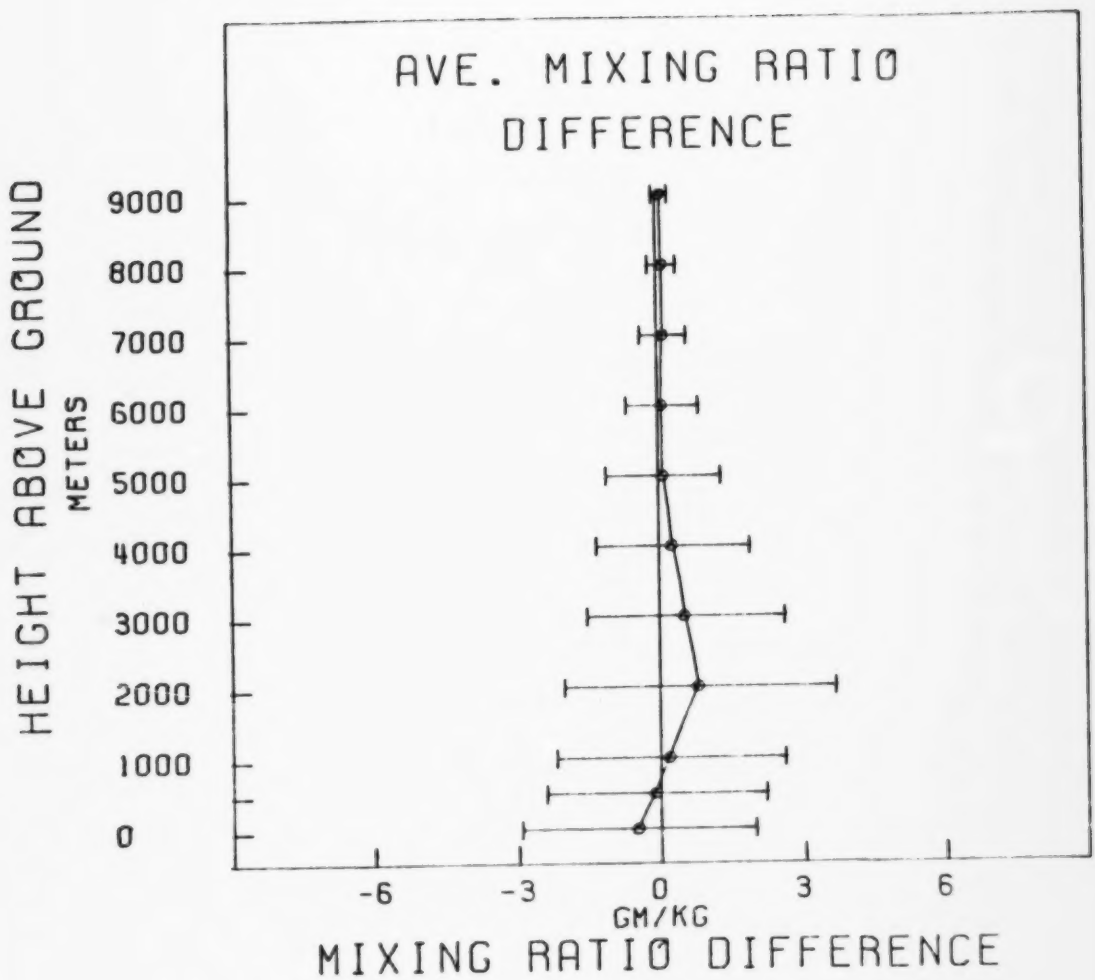

Figure 3.3 Average mixing ratio difference between the 115 pairs of tornado proximity soundings and associated check soundings.

Static energy and its use in connection with convective processes has been discussed by Riehl and Malkus (1958) and Darkow (1968). It is conservative with respect to saturated and unsaturated processes and it is directly proportional to pseudo-equivalent potential temperature or wet bulb potential temperature. Furthermore, it is easier to calculate than pseudo-equivalent potential temperature and can be used in the analysis of convective phenomenon in the same manner as pseudo-equivalent potential temperature.

Static energy can be approximated with negligible error by

$$E_s = C_p T + GZ + L_o W$$

where E_s is the static energy, C_p is the specific heat of air at constant pressure, T is the temperature, GZ is the geopotential, L_o is a constant value of the latent heat of condensation, and W is the mixing ratio (Darkow, 1968).

This equation was used to calculate values of static energy from the surface to 12 km (AGL) for all the soundings. The average values of static energy for the proximity and check soundings are shown in figure 3.4.

At corresponding heights above ground, the differences in static energy between the two average soundings are small. In contrast, Darkow (1969) found the proximity soundings to have substantially higher static energy values in the lower troposphere than the check soundings. The differing results of these two studies can be attributed to the different data sets used in the analyses. The data analyzed by Darkow (1969) contained roughly one half the number of cases used in this study. Also, not all cases in the earlier data set are contained within the data set analyzed here.

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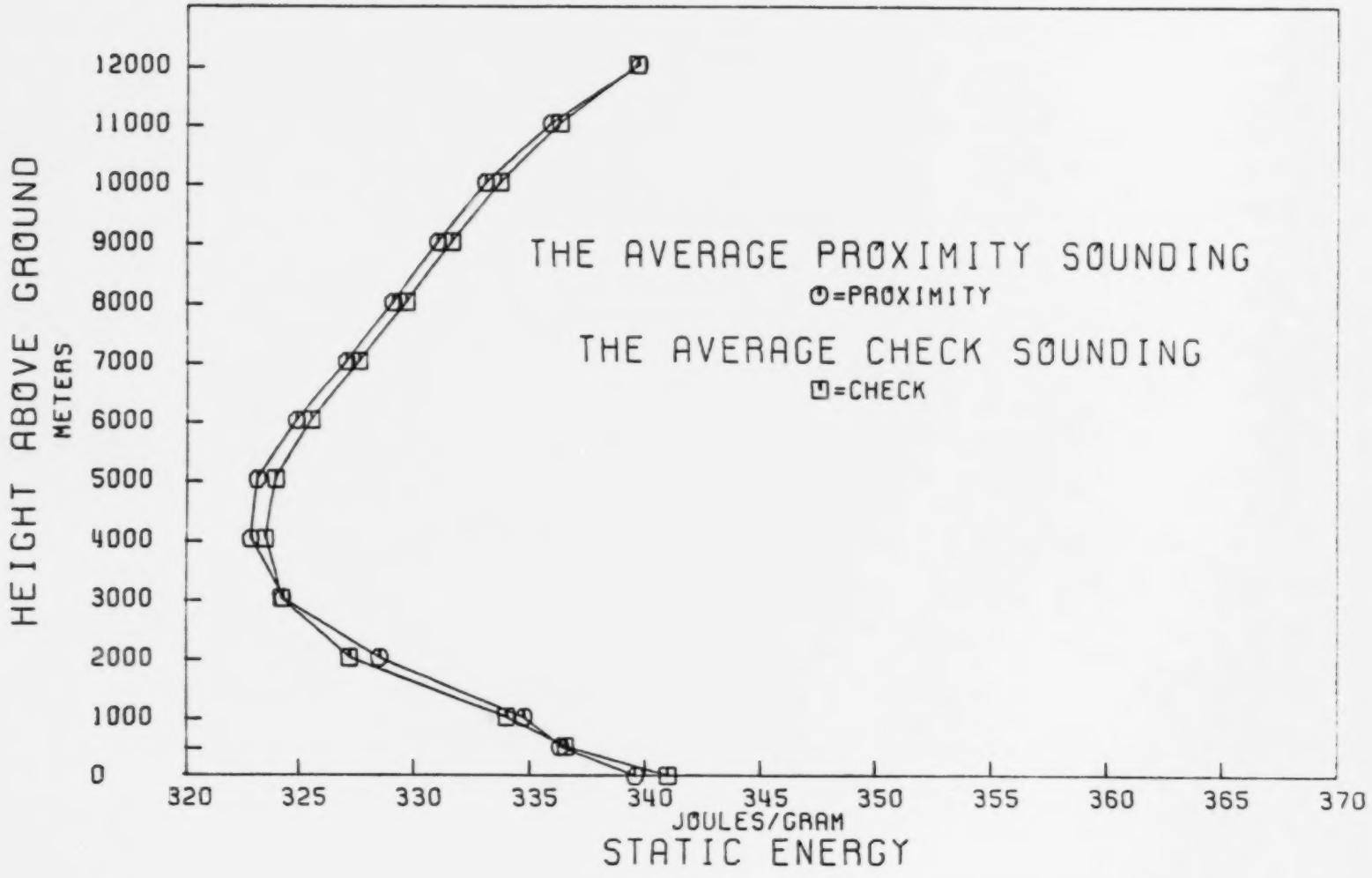

Figure 3.4 Average static energy profiles for the tornado proximity soundings and the associated check soundings (115 cases).

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The decrease of static energy with height is a measure of the potential-convective instability of the atmosphere. Viewed in this light, figure 3.4 indicates that both soundings have very similar degrees of potential-convective instability. The decrease in static energy from the surface to 4 km (AGL) is 17.5 joules/g for the mean check sounding and 16.7 joules/g for the mean proximity sounding. Therefore, the average check sounding actually has a slightly greater degree of potential-convective instability.

Figure 3.5 displays the average differences in static energy between the two sets of soundings. The average check sounding has slightly higher static energy values at all levels except in the layer from 1 to 3 km (AGL). For the levels where the average check sounding values are greater, the average difference is approximately 0.5 joules/g and is due primarily to the higher air temperatures of the check soundings. At 1 and 2 km (AGL), the static energy values of the proximity soundings averaged 1 joule/g greater than the check soundings. This is a result of the higher moisture content of the proximity soundings at these two levels. The standard deviations of the differences are quite large and overshadow the small average differences in static energy between the two sets of soundings.

The average wind speed profiles for the proximity and check soundings are shown in figure 3.6. It is readily apparent that the mean proximity sounding has greater wind speeds at all levels above the surface. The difference in wind speeds between the two average soundings maximizes at mid-tropospheric levels. In this region, the average proximity sounding's wind speeds are approximately 5 m/sec greater than the wind speeds of the average check sounding. The error inherent in wind speed

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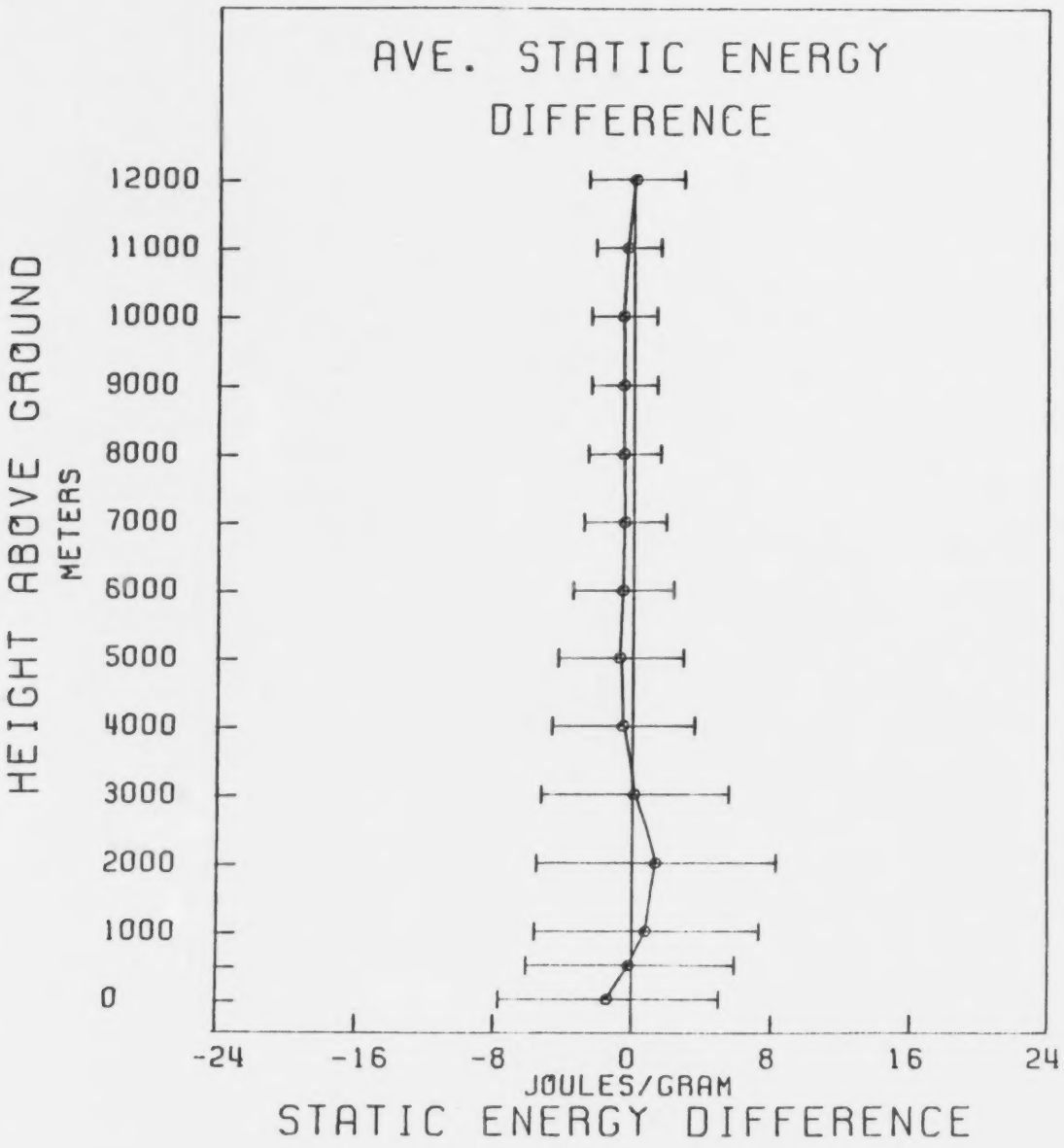

Figure 3.5 Average static energy difference between the 115 pairs of tornado proximity soundings and associated check soundings.

evaluations from rawinsonde data ranges from ± 1 m/sec at low levels to ± 5 m/sec at upper tropospheric levels (Kurihara, 1961).

It is important to note that both profiles show only moderate wind speeds. The average proximity sounding has winds of only 10 m/sec at 1 km (AGL) and only 25 m/sec at 12 km (AGL). Furthermore, in agreement with Maddox (1976), the variability of the wind speeds is quite large. Within the proximity sounding collection, wind speeds ranged from 1 to 32 m/sec at 1 km (AGL) and from 4 to 79 m/sec at 12 km (AGL). A frequency distribution of the wind speeds at 1, 5, and 12 km (AGL) for the proximity and check soundings is shown in table 3.1.

The layer of the atmosphere extending from the surface to 1 km (AGL) is characterized by strong shearing of the wind with height for both sets of soundings. Above this level, the average proximity sounding continues to exhibit strong shearing of the wind whereas the average check sounding does not. The average speed shear from 1 to 3 km (AGL) for the proximity soundings is 3.2 m/sec. However, the average check sounding shows a decrease in wind speed from 1 to 3 km (AGL); consequently, the average wind speed shear for the check sounding is -0.3 m/sec. The average difference in wind speed shear from 1 to 3 km (AGL) between the two sets of soundings is 3.4 m/sec. Shown in figure 3.7 is the relative frequency distribution of the difference in wind speed shear from 1 to 3 km (AGL) between the proximity soundings and check soundings. Values greater than zero indicate greater shear at the proximity station. The results indicate that in 77 percent of the cases the proximity soundings have greater wind speed shear in this layer than the associated check soundings. In an earlier study, Darkow and Fowler (1971) also noticed substantially greater wind speed shear in

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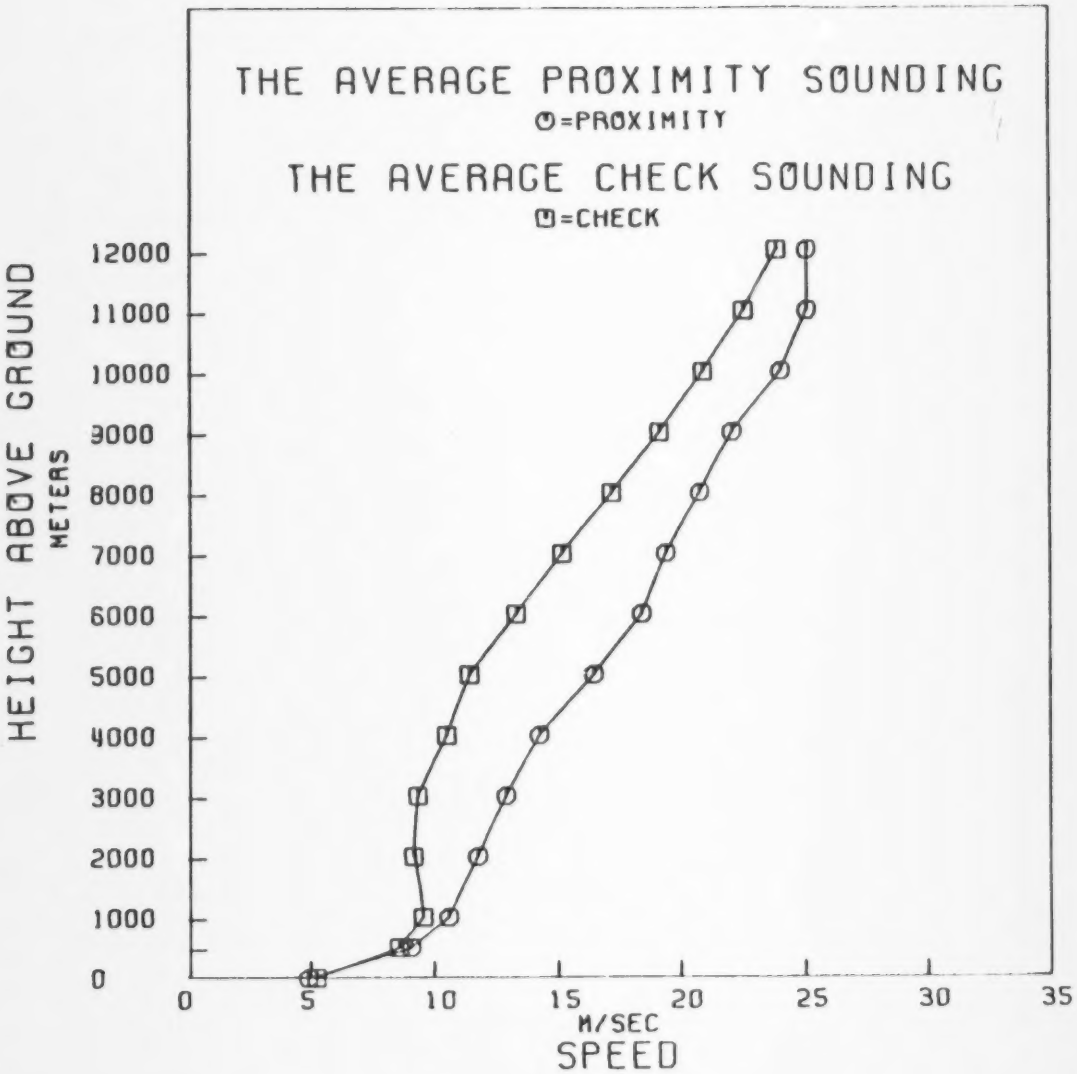

Figure 3.6 Average wind speed profiles for the tornado proximity soundings and the associated check soundings (115 cases).

TABLE 3.1

WIND SPEED DISTRIBUTION

FREQUENCY OF OCCURRENCE						
Type of Sounding	Class Intervals for Wind Speeds at 1 km (m/sec)					
	0 to 4.9	5 to 9.9	10 to 14.9	15 to 19.9	20 to 24.9	25 or Greater
Proximity	19	25	32	24	9	6
Check	13	40	29	20	7	6

FREQUENCY OF OCCURRENCE					
Type of Sounding	Class Intervals for Wind Speeds at 5 km (m/sec)				
	0 to 9.9	10 to 19.9	20 to 29.9	30 to 39.9	40 or Greater
Proximity	26	36	33	17	3
Check	47	37	27	2	2

FREQUENCY OF OCCURRENCE						
Type of Sounding	Class Intervals for Wind Speeds at 12 km (m/sec)					
	0 to 9.9	10 to 19.9	20 to 29.9	30 to 39.9	40 to 49.9	50 or Greater
Proximity	6	21	33	35	15	5
Check	6	25	40	24	15	5

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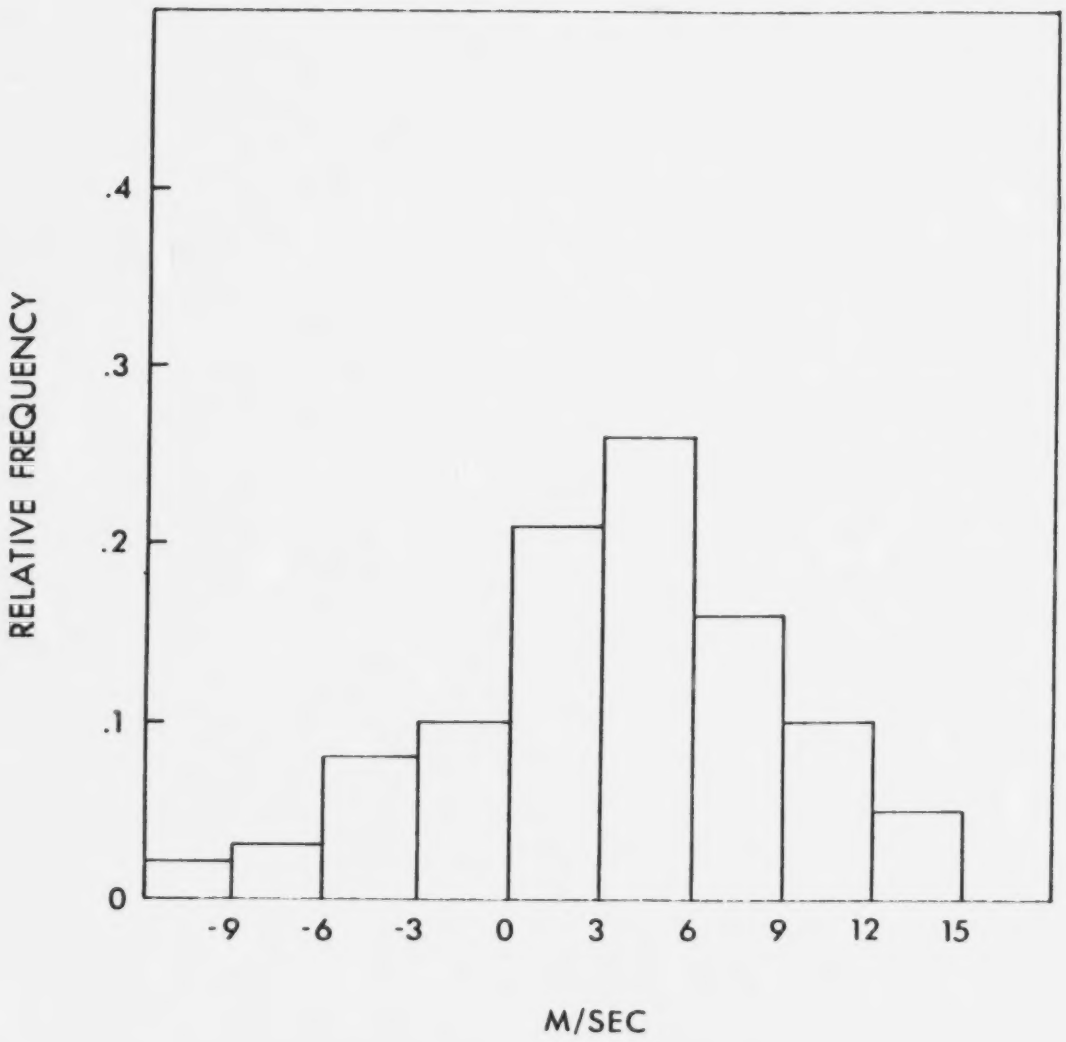

Figure 3.7 Relative frequency distribution of the difference in wind speed shear in the 1 to 3 km layer of the atmosphere between the proximity soundings and the check soundings (115 cases).

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the proximity soundings as compared to the check soundings. Gray (1969) also mentions the importance of lower tropospheric wind shear to tornado genesis. Barnes (1970) suggests that large amounts of localized vorticity about a vertical axis comparable to the cyclonic vorticity of a typical tornado can be generated when a strongly sheared layer encounters the differential vertical velocity field of a thunderstorm. Therefore, the wind speed shear from 1 to 3 km (AGL) may be a good tool for delineating the location of possible severe weather development.

By analyzing tornado proximity data, Wills (1969) and Gray (1971) found that the vector shear from the surface to mid-tropospheric levels averaged 20 to 25 m/sec in severe storm environments. Similar results were obtained from this data set. The average vector shear from the surface to 5 km (AGL) is 19.5 m/sec for the proximity sounding collection. The corresponding average vector shear for the check soundings is only 14.3 m/sec. The resulting average difference in vector shear between the two sets of soundings is 5.2 m/sec. The relative frequency distribution of the difference in vector shear between the two sets of soundings (not shown) is very similar to the corresponding distribution for wind speed shear from 1 to 3 km (AGL). In fact, results are identical in one aspect. The vector shear of the proximity soundings was greater than the vector shear of the associated check soundings in 77 percent of the cases. Therefore, the vector shear from the surface to 5 km (AGL) appears to provide information similar to that obtained from the speed shear in the 1 to 3 km (AGL) layer.

Another variable of interest is the horizontal moisture flux. This is simply the product of the wind speed and the absolute humidity. Physically, it is the amount of moisture passing through a unit vertical

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plane perpendicular to the wind per unit time. Its importance arises from the fact that it gives an estimate of the rate at which fuel might be supplied to a stationary convective storm. The estimate is only approximate since storm movement is not considered in its calculation.

The average values of moisture flux for the lowest 2 km of the atmosphere for the proximity and check soundings are 123 and 119 g/m^2sec, respectively. Also, the wind directions in the lowest 2 km of the atmosphere of the average check and average proximity soundings are quite similar (see figure 3.8). Thus, the environmental conditions at both the proximity station and the check station are capable of providing fuel to a convective storm at essentially the same rate.

It is important to note that the moisture flux values are characterized by extreme variability. Values as low as 15 g/m^2sec and as high as 340 g/m^2sec were observed within the proximity sounding collection. Similar extreme values were noted within the check sounding collection.

Figure 3.8 shows the average proximity sounding hodograph and the average check sounding hodograph. The winds at corresponding heights from the surface to 1 km (AGL) are similar for the two hodographs. Both hodographs have a south-southeast surface wind (170^o) which veers and increases in magnitude with height resulting in a south-southwest wind (195^o) at 1 km (AGL). Above this level, both hodographs continue to exhibit veering of the wind with height. The amount of veering in the 1 to 5 km (AGL) layer is 51^o for the average check sounding and 43^o for the average proximity sounding. This result tends to discredit the utility of the veering term in the SWEAT Index (Miller and Maddox, 1975).

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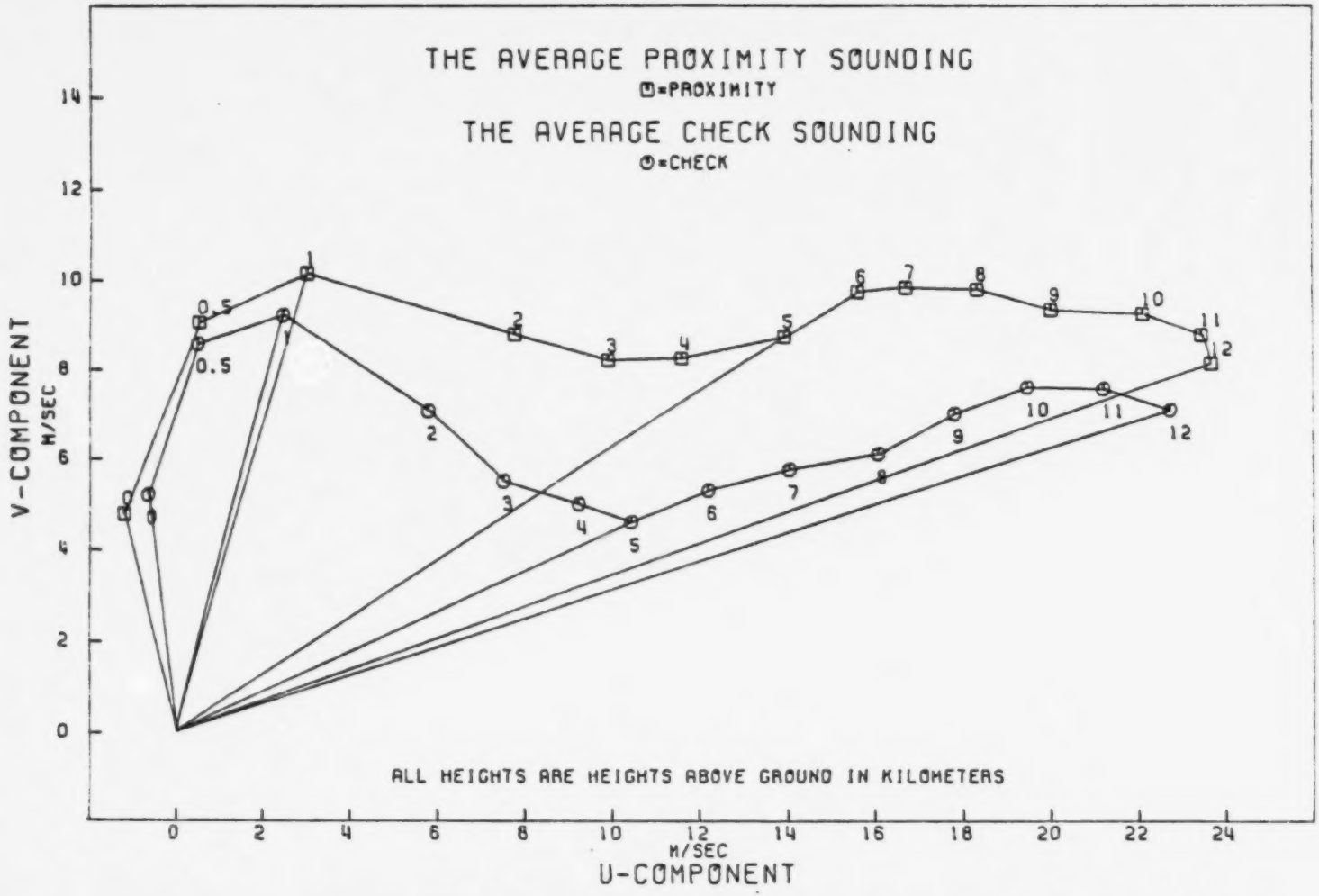

Figure 3.8 Average hodographs for the tornado proximity soundings and the associated check soundings (115 cases).

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Although the veering of the wind in the 1 to 5 km (AGL) layer is similar for the two average soundings, the wind speed increases within this layer are not similar. The average check soundings winds increase in speed by only 2 m/sec in this layer; the corresponding increase in wind speed for the average proximity sounding is 6 m/sec. As noted earlier, the mid-troposphere is the region of maximum difference in wind speed between the two sets of soundings.

The veering of the wind in the 5 to 12 km (AGL) layer is such that the two average soundings have nearly identical wind directions at 12 km (AGL). Furthermore, the average check sounding's winds increase in speed more than the average proximity sounding's winds within this layer resulting in very similar wind speeds at 12 km (AGL).

In summary, the wind directions at corresponding levels are similar for the two average soundings. The maximum difference in wind direction, which occurs in mid-tropospheric levels, is only 8^{o}. However, as noted earlier, there are substantial differences in wind speed and in wind speed shear between the two sets of soundings.

2. Analysis of the 12 Hour Change Prior to the Tornado Proximity Sounding

The temperature and dew points of the average proximity sounding and the average 12 hour precedent sounding are shown in figure 3.9. At 2 km (AGL) and above the temperature profiles are nearly identical. From the surface to 2 km (AGL) the temperature profiles reflect the expected diurnal temperature change. The average 12 hour precedent sounding has a nearly isothermal layer from the surface to 0.5 km (AGL) and an average lapse rate of 4.6^{o}C per km in the layer extending from 0.5 km (AGL) to 2 km (AGL). By late afternoon, the

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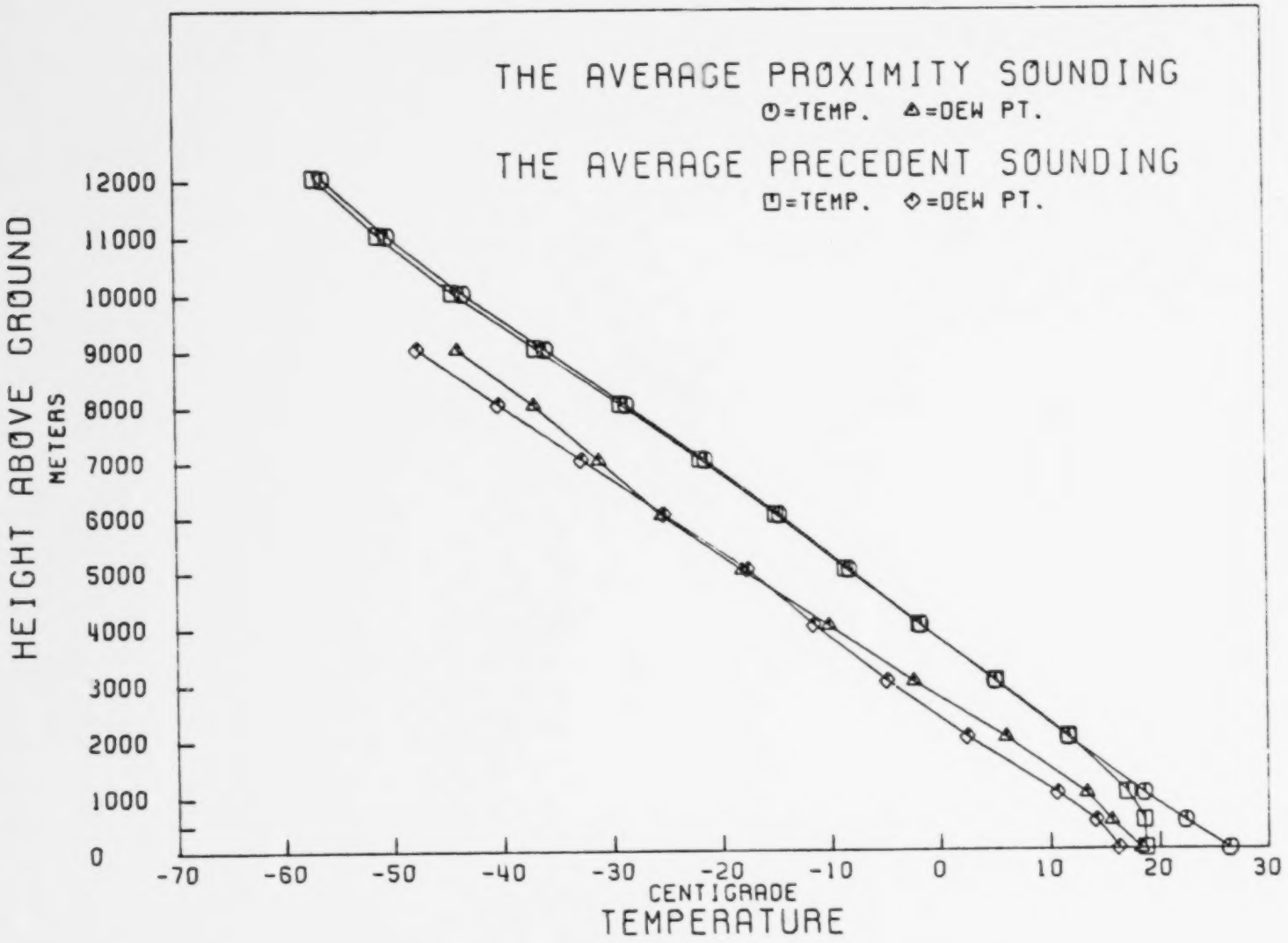

Figure 3.9 Average temperatures and dew point temperatures for the tornado proximity soundings and the associated 12 hour precedent soundings (115 cases).

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lapse rate in the lowest 2 km of the atmosphere has increased to 7.5°C per km. The increased lapse rate is a result of the substantial diurnal increases in temperature at 1 km (AGL) and below which were produced by surface heating and small scale convective mixing. The maximum temperature increase occurs at the surface. Here, the average temperature increased from approximately 18°C to approximately 26°C during the 12 hour period. Smaller increases in temperature are noted at 0.5 and 1.0 km (AGL).

The most pronounced change in the dew point temperatures occurs below 4 km (AGL). The average surface dew point temperature increased from its morning value of 16°C to 19°C by late afternoon. Similar daytime increases in dew point temperature occur through 3 km (AGL). These results are very similar to those found by Lustig (1973) in an earlier analysis of proximity and precedent sounding data.

It is important to mention that these profiles are only an average, and that conditions in which severe weather develops are extremely variable. For example, surface temperatures ranged from 11°C to 38°C among the proximity sounding data. Also, surface values of mixing ratio as low as 8 g/kg and as high as 22 g/kg were observed within the proximity sounding collection.

Figure 3.10 shows the average temperature increase and its variability from the morning (1200 GMT) to the late afternoon (0000 GMT). It is readily apparent that the most pronounced average diurnal temperature increases occur below 2 km (AGL). In contrast, the results indicate that on the average, there was no daytime temperature changes at 2 and 3 km (AGL). Above 3 km (AGL), the average temperature increases are 1°C or less and the variabilities are such that the

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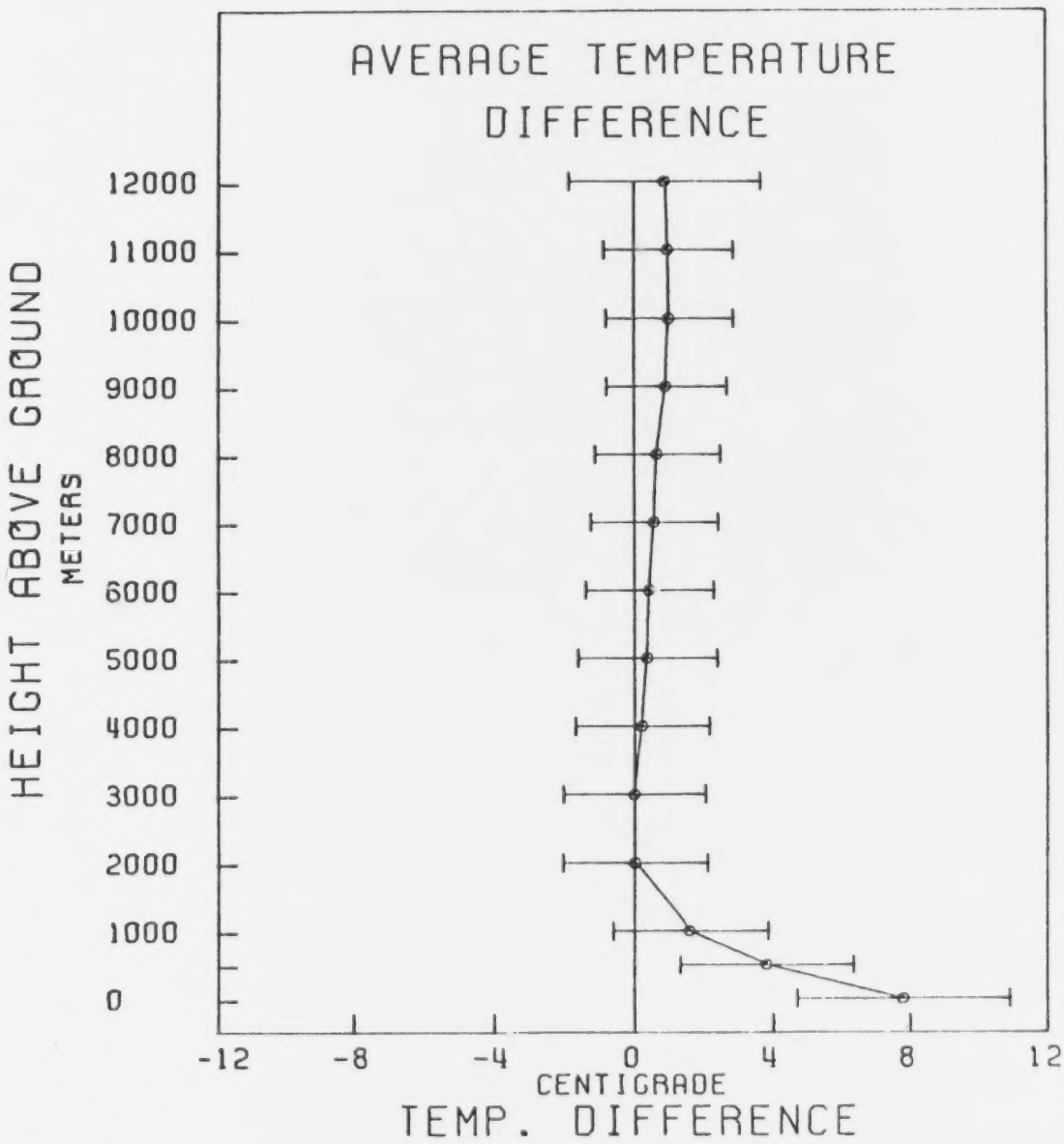

Figure 3.10 Average temperature difference between the 115 pairs of tornado proximity soundings and associated 12 hour precedent soundings.

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probability of a diurnal temperature increase ranges from 55 percent at 4 km (AGL) to 70 percent at 11 km (AGL).

Whitney and Miller (1956) determined that differential temperature advection was the destabilization mechanism in one particular severe weather incident. Since cooling at mid-tropospheric levels occurred in approximately 45 percent of the cases in this data set, differential temperature advection might have been a destabilization mechanism in many instances. In order to acquire more information about this hypothesis, the horizontal temperature advection was calculated using the technique described by Panofsky (1956). Both the morning and evening winds were included in this computation to establish an average daytime horizontal temperature advection. The results indicate that above the boundary layer the average rate of temperature advection is approximately + 0.11^{o}C/hour for all the 1 km thick layers from 2 to 12 km (AGL) except the layer from 5 to 6 km (AGL). The average rate of horizontal temperature advection in this layer is less than 0.01^{o}C/hour. Thus, daytime cooling in this layer due to advection occurred in nearly 50 percent of the cases. Furthermore, daytime cooling in adjacent layers due to advection occurred in approximately 40 percent of the cases. This result adds credibility to the hypothesis that differential temperature advection may have been a destabilization mechanism in many of these severe weather events. Also, the magnitudes of the average horizontal temperature advection are sufficient to account for the average daytime warming in all the 1 km thick layers above the boundary layer except for the layer from 5 to 6 km (AGL).

Figure 3.11 displays the average diurnal mixing ratio increases or decreases and their variabilities. The figure shows a substantial

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daytime increase in moisture below 4 km (AGL). The increase in moisture from the surface to 3 km (AGL) averaged 1.5 g/kg which is nearly identical to the average moisture increase found by Lustig (1973) for the same portion of the atmosphere. Furthermore, this water vapor enrichment corresponds to an increase in moisture of approximately 0.5 g/cm^2. Measurements, cited by Steycart (1973), from a lysimeter operated by the University of Missouri at Columbia indicate that this degree of moisture enrichment may be accomplished solely by local evapotranspiration from a typical vegetative cover with ample soil moisture. However, there is no question that in many instances the advection of moisture was very important and was partially responsible for the diurnal moisture increase. Above 3 km (AGL) the average diurnal moisture changes are quite small, less than 0.3 g/kg, and they do not substantially alter the stability of the atmosphere.

One of the most interesting features of this figure is the moisture increase at 0.5 km (AGL). It is only 60 percent as large as the moisture increase at the surface and at 1 km (AGL). A possible explanation for this rather surprising result is the presence of moisture inversions in a number of the precedent (1200 GMT) soundings. For example, a precedent sounding released at Amarillo, Texas on June 21, 1976, had a mixing ratio of 8 g/kg at the surface and 10 g/kg at 0.5 km (AGL). By evening, the moisture structure of the lower boundary layer had been modified such that the mixing ratios at the surface and at 0.5 km (AGL) were 10 g/kg and 9 g/kg, respectively. In contrast, the moisture content of the atmosphere at 1 km (AGL) remained essentially unchanged throughout the day at 8 g/kg. The daytime modifications transformed the moisture structure of the atmosphere from a

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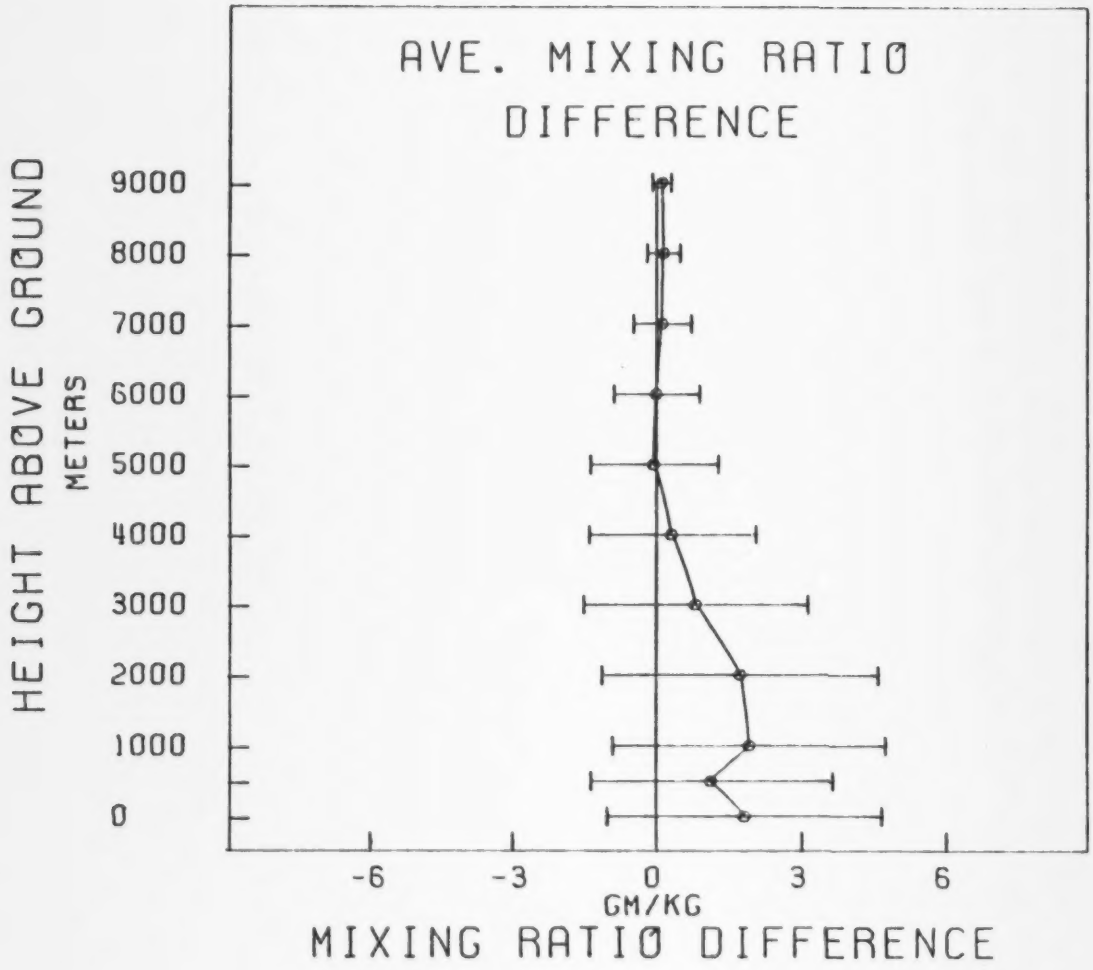

Figure 3.11 Average mixing ratio difference between the 115 pairs of tornado proximity soundings and associated 12 hour precedent soundings.

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moisture inversion condition to a moisture lapse condition. In the process, there was a moisture addition of 2 g/kg at the surface, a moisture decrease of 1 g/kg at 0.5 km (AGL), and a negligible moisture change at 1 km (AGL). A number of cases similar to this example could account for the smaller diurnal moisture increase at 0.5 km (AGL).

The average static energy profiles for the proximity and precedent soundings are shown in figure 3.12. The average proximity sounding has higher static energy values at all levels. The average daytime increases in static energy above 4 km (AGL) are small and are a result of the average diurnal increases in temperature at these levels. Below 4 km (AGL), the average diurnal increases in static energy are quite large. Tnese low level increases are a result of the diurnal increases in temperature and moisture at these levels.

The decrease of static energy from its maximum value in the boundary layer to 4 km (AGL) is 16.7 joules/g for the average proximity sounding and 7.9 joules/g for the average precedent sounding. Thus, the average proximity sounding possesses significantly more potential-convective instability as compared to the average precedent sounding. This result clearly demonstrates the importance of diurnal changes in temperature and moisture upon the stability of the atmosphere.

Figure 3.13 shows the average diurnal increase in static energy and its variability from the surface to 12 km (AGL). The largest increase, 12.3 joules/g, occurs at the surface. The average daytime increase in static energy lessens with increasing height to 5 km (AGL) where the minimum increase in static energy, 0.3 joules/g, occurs. In the layer from 6 to 12 km (AGL), the average diurnal increase in static energy ranges from a minimum of 0.5 joules/g at 6 km (AGL) to a

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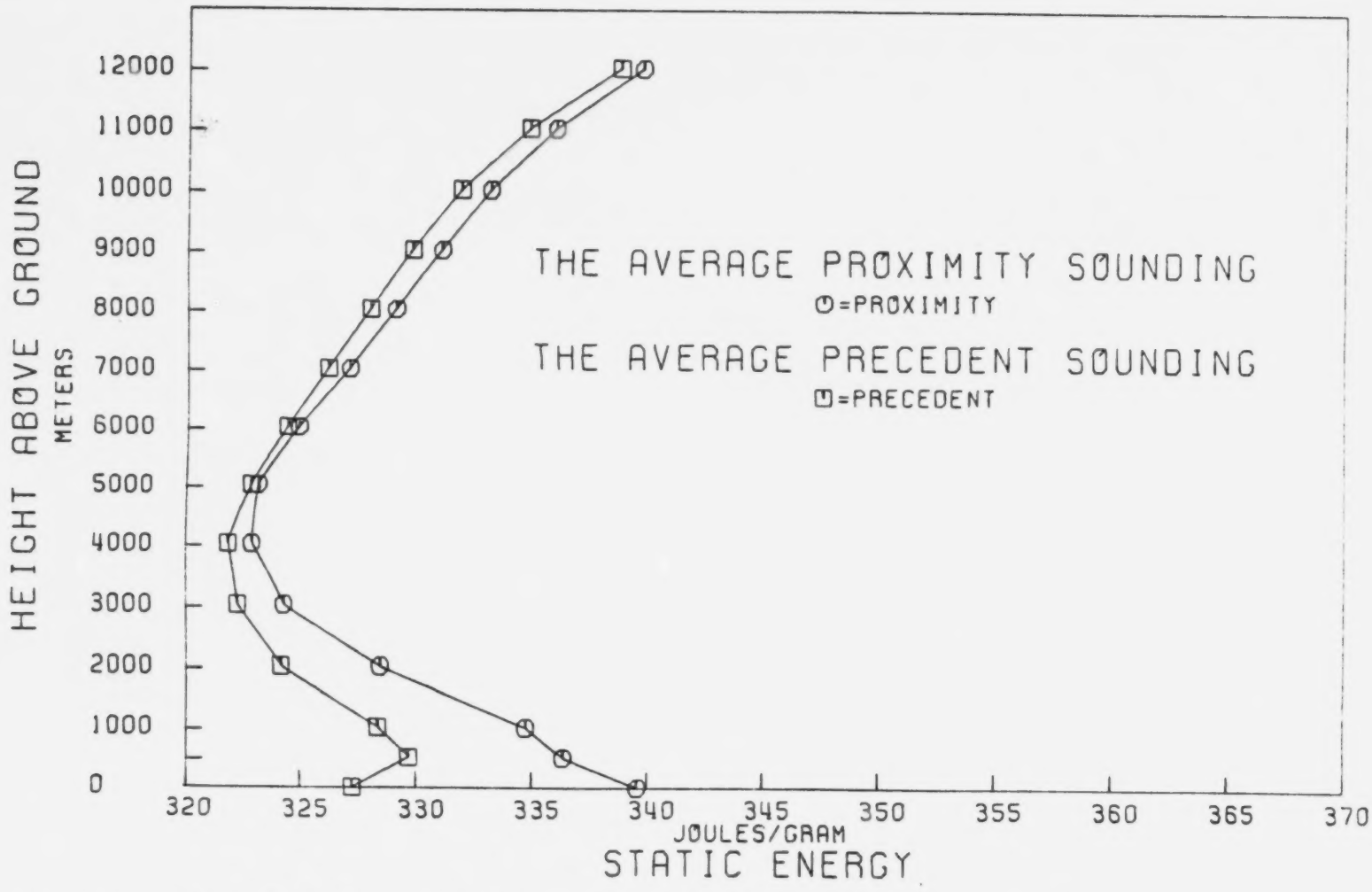

Figure 3.12 Average static energy profiles for the tornado proximity soundings and the associated 12 hour precedent soundings (115 cases).

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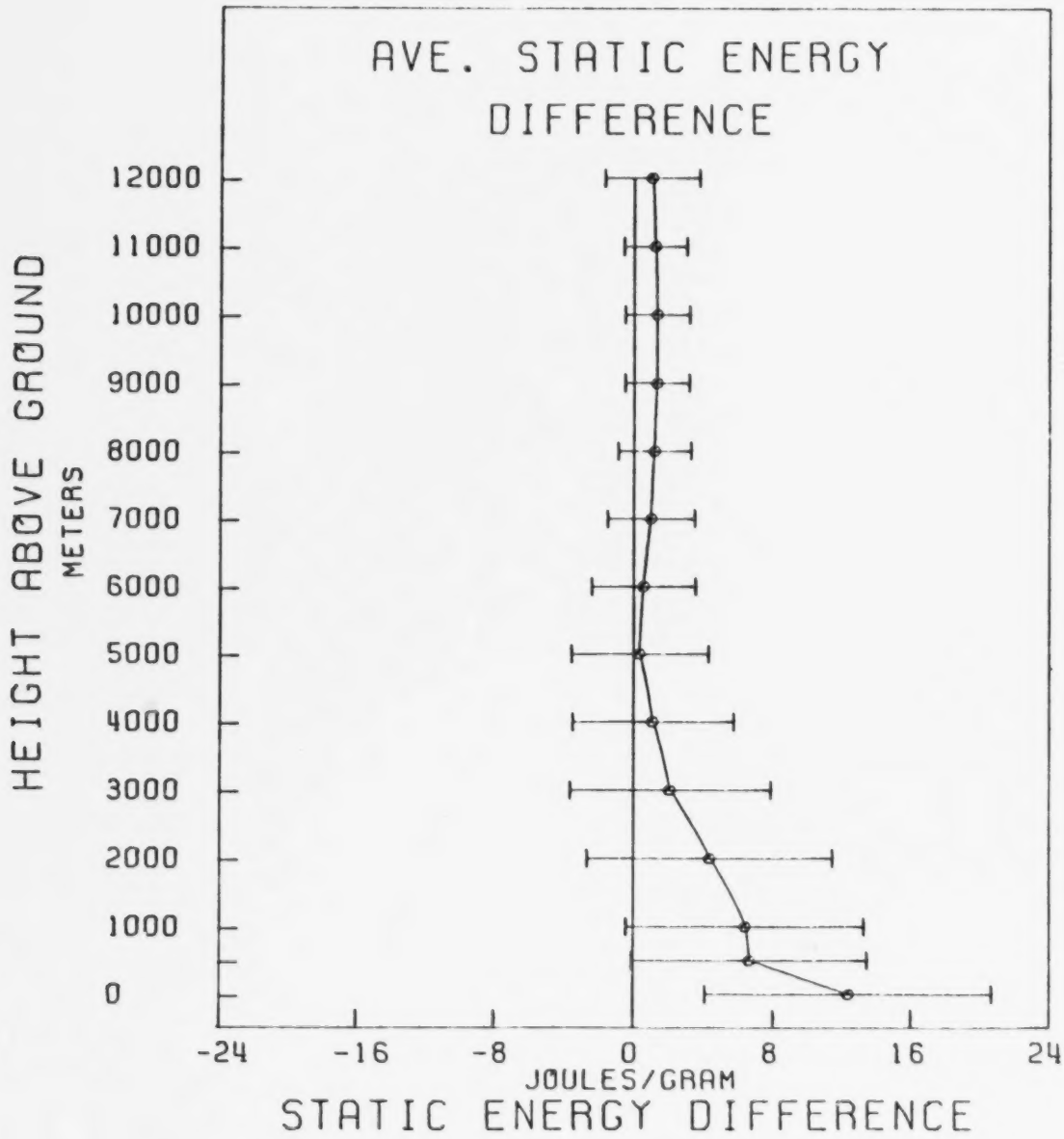

Figure 3.13 Average static energy difference between the 115 pairs of tornado proximity soundings and associated 12 hour precedent soundings.

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maximum of 1.2 joules/g at 10 km (AGL). However, due to the magnitude of the increases and their impact upon instability, the most important diurnal increases in static energy are those that transpired below 4 km (AGL).

The average wind speed profiles for the two sets of soundings are shown in figure 3.14. Although both profiles show strong positive shear in the lowest kilometer of the atmosphere, above 1 km (AGL) the two profiles exhibit some substantial differences in speed shear.

The average precedent sounding's wind speed decreases from 1 to 3 km (AGL) resulting in a jet-like profile. In contrast, the wind speed profile for the average proximity sounding exhibits strong positive shear in this layer. The average speed shear from 1 to 3 km (AGL) is 3.2 m/sec for the proximity soundings. For the precedent soundings, the average speed shear in the corresponding layer is -0.7 m/sec. The average difference in speed shear between the two sets of soundings is 3.9 m/sec. Examination of the frequency distribution of the difference in wind speed shear between the two sets of soundings indicates that in 75 percent of the cases the proximity sounding has greater wind speed shear from 1 to 3 km (AGL) than the associated precedent sounding. However, above 3 km (AGL) the wind speed shear is similar for both profiles.

The average proximity sounding has stronger winds at all levels above 1 km (AGL) as compared to the average precedent sounding. The average wind speed difference between the two sets of soundings above 1 km (AGL) is approximately 3 m/sec. The region of maximum wind speed difference is from 5 to 8 km (AGL) where the average proximity sounding's wind speed is approximately 4 m/sec stronger than the average precedent

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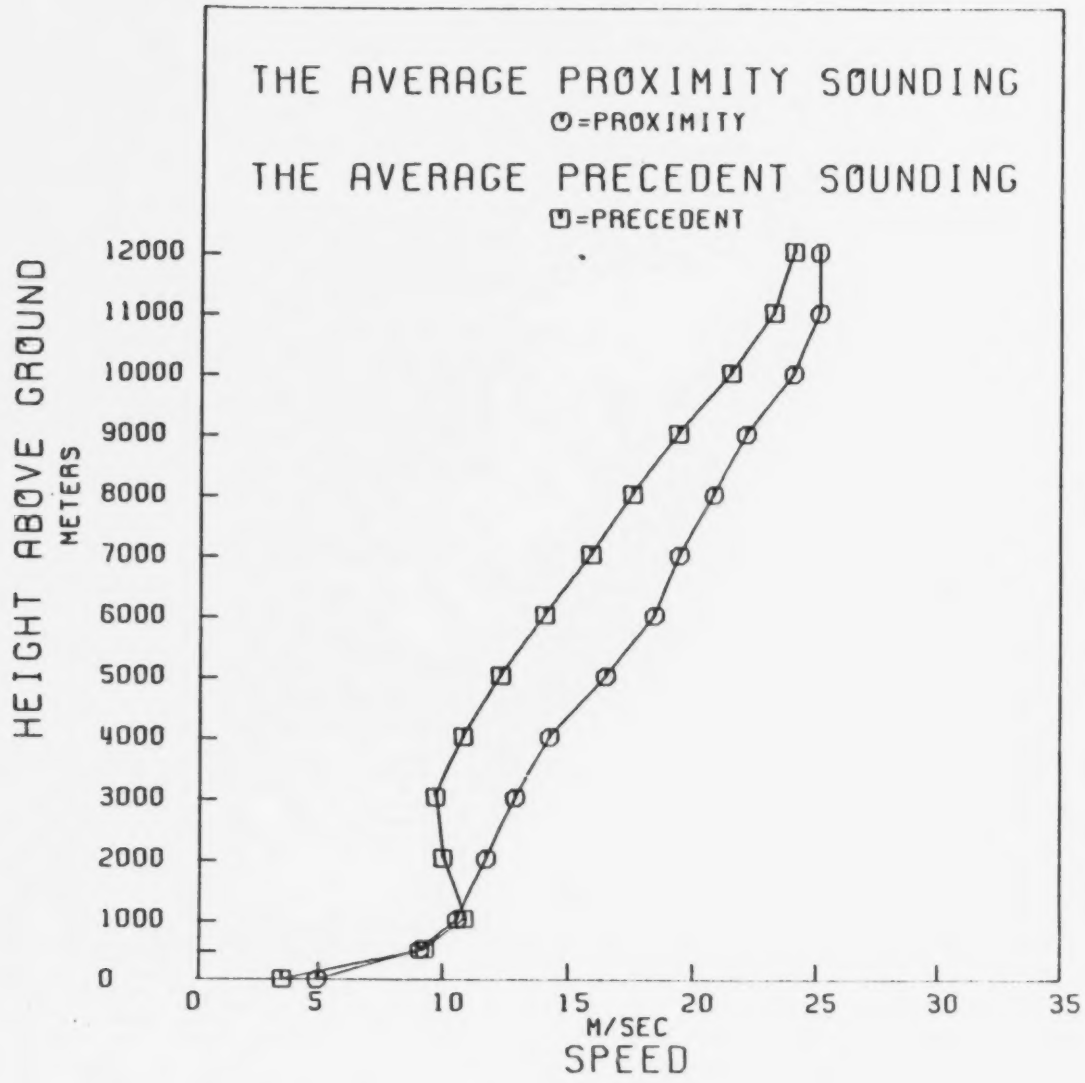

Figure 3.14 Average wind speed profiles for the tornado proximity soundings and the associated 12 hour precedent soundings (115 cases).

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TABLE 3.2

WIND SPEED DISTRIBUTION

FREQUENCY OF OCCURRENCE						
Type of Sounding	Class Intervals for Wind Speed at 1 km (m/sec)					
	0 to 4.9	5 to 9.9	10 to 14.9	15 to 19.9	20 to 24.9	25 or Greater
Proximity	19	25	32	24	9	6
Precedent	13	33	29	19	15	6

FREQUENCY OF OCCURRENCE					
Type of Sounding	Class Intervals for Wind Speed at 5 km (m/sec)				
	0 to 9.9	10 to 19.9	20 to 29.9	30 to 39.9	40 or Greater
Proximity	26	36	33	17	3
Precedent	27	55	31	2	0

FREQUENCY OF OCCURRENCE						
Type of Sounding	Class Intervals for Wind Speeds at 12 km (m/sec)					
	0 to 9.9	10 to 19.9	20 to 29.9	30 to 39.9	40 to 49.9	50 or Greater
Proximity	6	21	33	35	15	5
Precedent	6	23	39	26	16	5

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sounding's winds. A frequency distribution of the wind speeds at 1, 5 and 12 km (AGL) for the proximity and precedent soundings is shown in table 3.2.

These results indicate that there is a substantial diurnal increase in wind speed above 1 km (AGL) in the region of severe weather development. Furthermore, one of the more important consequences of the diurnal wind speed increases is the transformation of a jet-like wind speed profile to one of strong positive shear throughout.

Referring to figure 3.6, it is interesting to note the striking similarity between the average check sounding wind speed profile and the average precedent sounding wind speed profile. Both profiles have a jet-like structure with a relative wind speed maximum at 1 km (AGL). Also, above 3 km (AGL) both profiles have similar wind speeds at corresponding levels and similar increases in wind speed at corresponding height increments.

As mentioned earlier, the average value of moisture flux in the lowest 2 km of the atmosphere for the proximity soundings is 123 $g/m^2 sec$. The average value in the same layer for the precedent soundings is 105 $g/m^2 sec$. This amounts to a 17 percent diurnal increase in the relative rate at which the environment can supply fuel to a convective storm. However, since the corresponding average wind directions in the lowest 2 km of the atmosphere are considerably different between the two sets of soundings (see figure 3.15), the actual diurnal increase in the rate of fuel supply to a convective storm would depend upon storm movement.

The 17 percent increase in moisture flux results from the daytime increase in low level moisture. However, this daytime increase is

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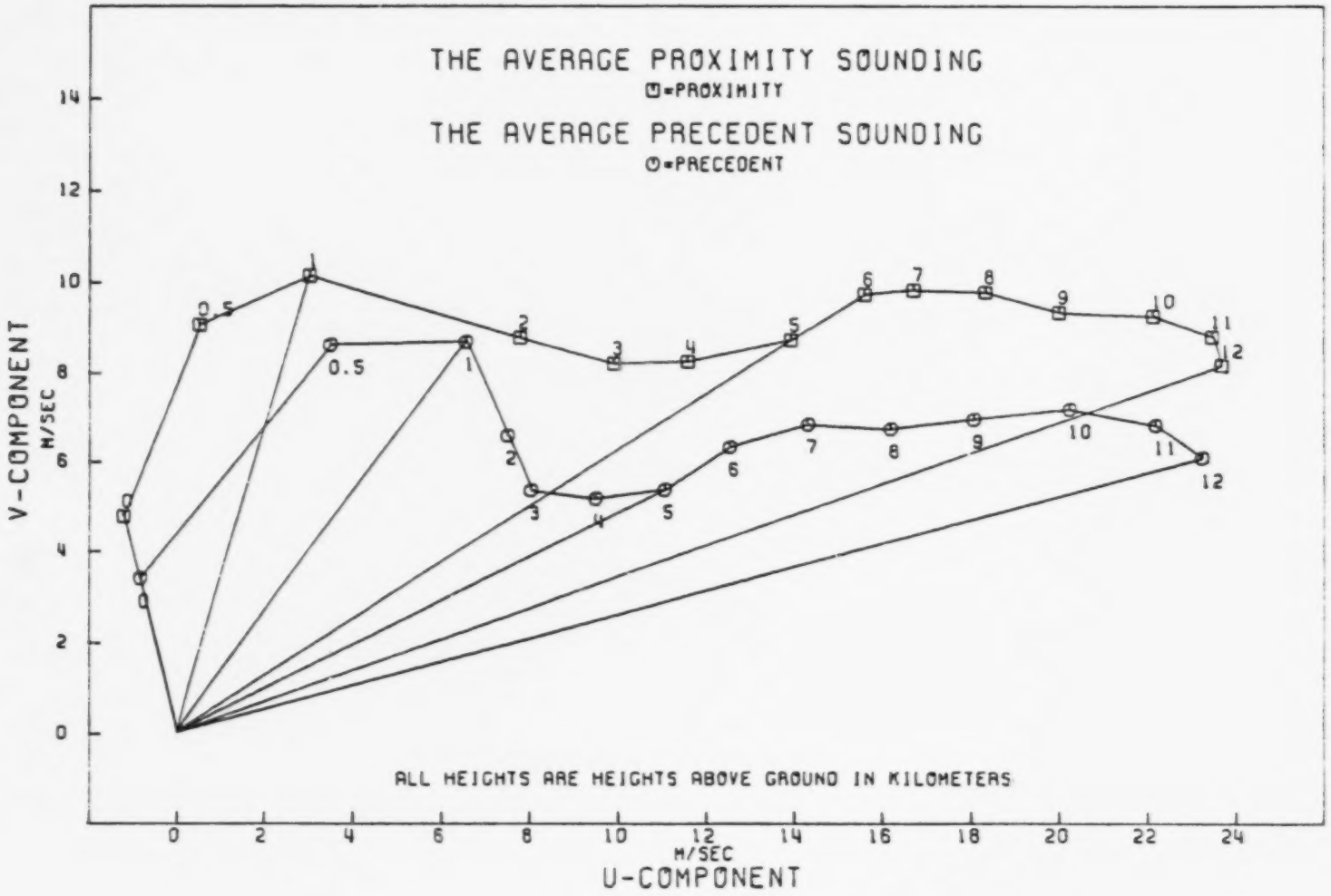

Figure 3.15 Average hodographs for the tornado proximity soundings and the associated 12 hour precedent soundings (115 cases).

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overshadowed by the wide range of moisture flux values among the two sets of soundings. Values as low as 15 g/m^2sec and as high as 340 g/m^2sec were observed within this collection.

Figure 3.15 shows the hodographs for the average precedent and the average proximity sounding. At the surface the wind directions are identical; however, the figure shows a noticeable daytime increase in wind speed at this level. Above the surface, the average precedent sounding exhibits substantially more veering of the wind with height to 1 km (AGL) than the average proximity sounding. Thus, the diurnal change in wind velocity at 0.5 and 1 km (AGL) is characterized by small speed changes, but significant backing of the wind.

Above 1 km (AGL), the diurnal change in wind direction continues to be characterized by a shift to a more southerly direction. Although the backing of the wind is rather small (4^o to 7^o) at these levels, there is a substantial diurnal increase in wind speed at all levels above 1 km (AGL).

It is hypothesized that the diurnal wind velocity changes are in response to the eastward migration of the approaching mid-level trough.

3. Differences in Daytime Destabilization

Displayed in Figure 3.16 are the average differences between the 12 hour temperature changes at the proximity and check sounding stations. The tick mark equidistant on either side of the average values represent plus or minus one standard deviation. Positive values indicate a greater diurnal increase in temperature at the proximity station. Alternatively, negative values indicate a greater daytime

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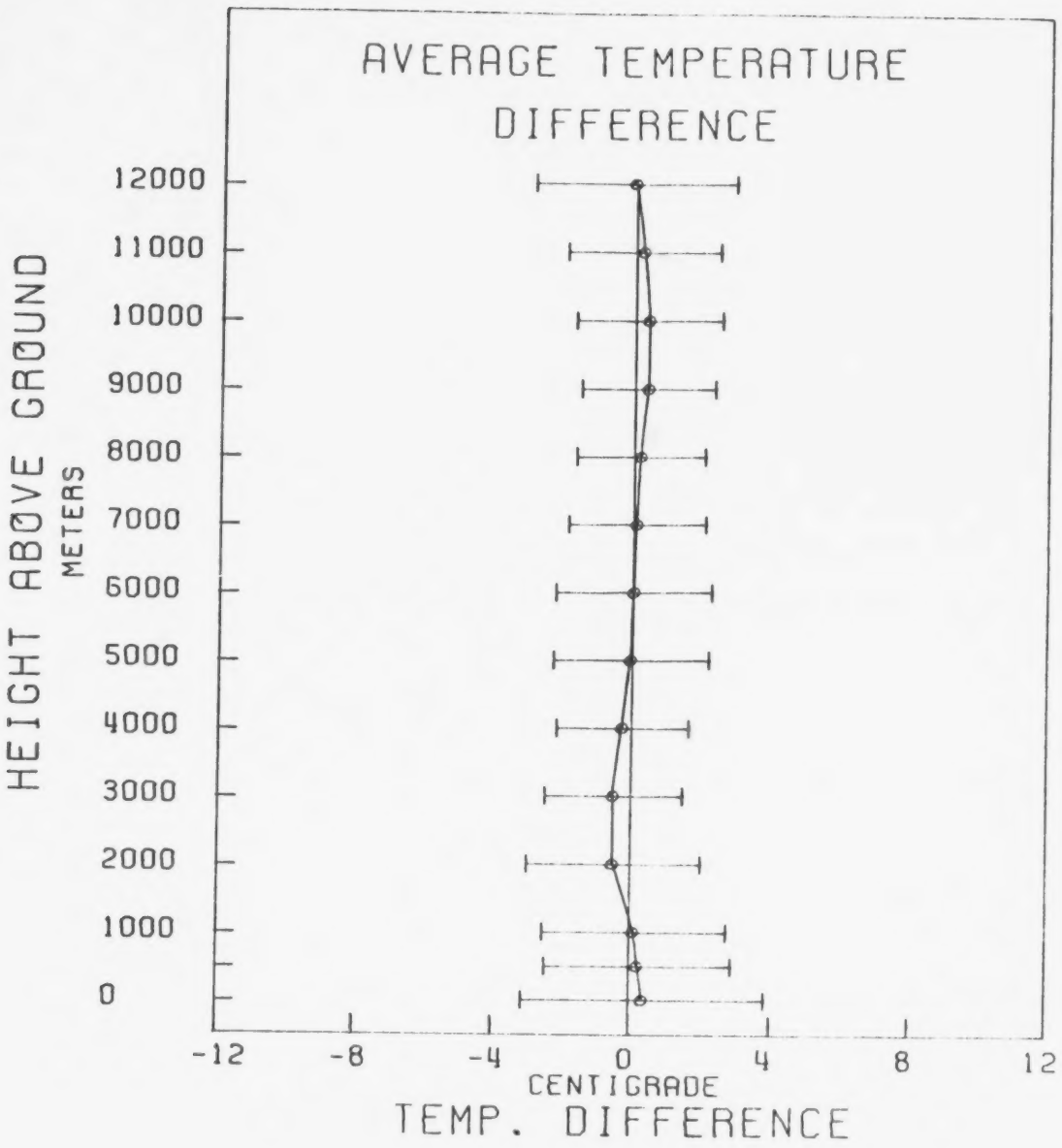

Figure 3.16 Average difference between the 12 hour temperature changes at the proximity station and at the check station (115 cases).

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increase in temperature at the check station. Figures 3.17 and 3.18 will have the same type format.

From figure 3.16, it is readily apparent that the average differences in diurnal temperature changes are very small. The average difference of greatest magnitude, $-0.5^{o}C$, occurs at 2 and 3 km (AGL) and indicates slightly greater diurnal warming at the check station. The proximity station experiences greater daytime warming below 2 km (AGL) and in the layer from 3 to 11 km (AGL). The layer from 5 to 7 km (AGL) experiences virtually no average difference in diurnal temperature change at the two locations.

Because the average differences are very small and are not significant, the results suggest that on the average both locations experience the same degree of diurnal temperature increase at all levels from the surface to 12 km (AGL). However, since the variability is large, any one case may show substantial differences in the daytime temperature changes between the proximity and check soundings stations.

The average differences and variabilities in the 12 hour moisture changes at the two locations are shown in figure 3.17. At all levels, the average daytime increase in moisture at the proximity station is equal to or greater than the average increase in moisture at the check station. Above 2 km (AGL) the difference is small, 0.3 g/kg or less, and is insignificant. However, the average moisture increase at the proximity station from the surface to 2 km (AGL) is 0.8 g/kg larger than the corresponding average increase in moisture at the check station. The variabilities suggest that in 60 percent of the cases the proximity sounding station experienced a greater diurnal increase in moisture at 2 km (AGL) and below than the associated check sounding station. Recall

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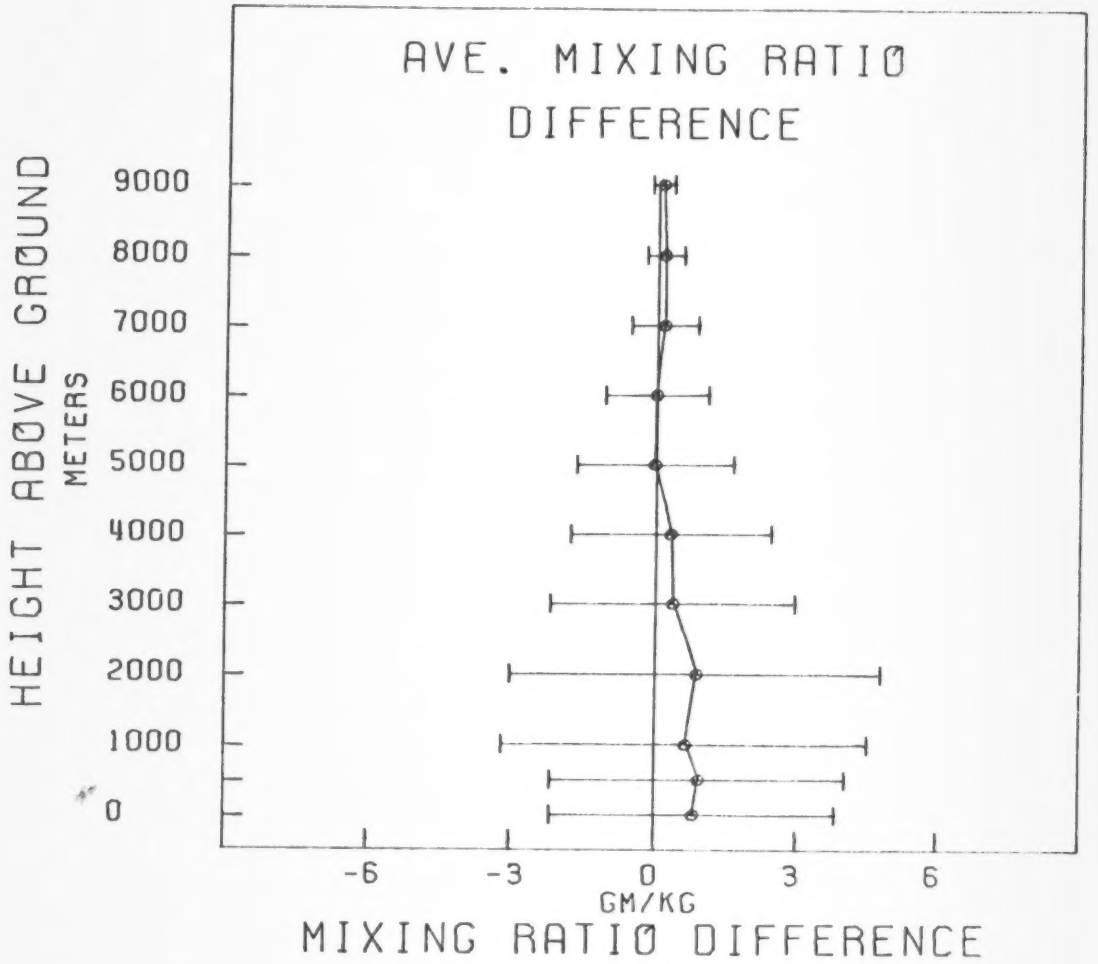

Figure 3.17 Average difference between the 12 hour mixing ratio changes at the proximity station and at the check station (115 pairs).

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that the average diurnal increase in moisture in this layer at the proximity sounding station was 1.5 g/kg; therefore, the average moisture increase in the near vicinity of the severe weather development was twice the average increase in moisture at the closest upper air station in the upstream direction. The difference in diurnal moisture increases may have been due to stronger local evapotranspiration and/or stronger horizontal moisture converegence at the proximity sounding station. Endlich and Mancuso (1968), Hudson (1971) and others have commented on the importance of moisture converegence to severe storm occurrences.

Figure 3.18 displays the average differences and variabilities in the 12 hour static energy increases at the proximity sounding and check sounding stations. The diurnal increase in static energy at the proximity station is equal to or greater than the corresponding increase at the check station at all levels. Above 2 km (AGL), the maximum average difference is only 0.6 joules/g. However, at 2 km (AGL) and below the impact of the differences in low level moisture addition between the two stations is readily apparent. The proximity sounding station's diurnal increase in static energy averages 2 joules/g greater than the corresponding increase at the check sounding station at these levels. Furthermore, the results indicate that in 60 percent of the cases the diurnal increase in static energy at 2 km (AGL) and below at the proximity sounding station is greater than the increase in static energy at the associated check sounding station.

The results above indicating a greater diurnal increase in low level moisture and static energy at the proximity sounding as compared to the check sounding station must be tempered by the comparison of proximity soundings versus check soundings discussed in section 1

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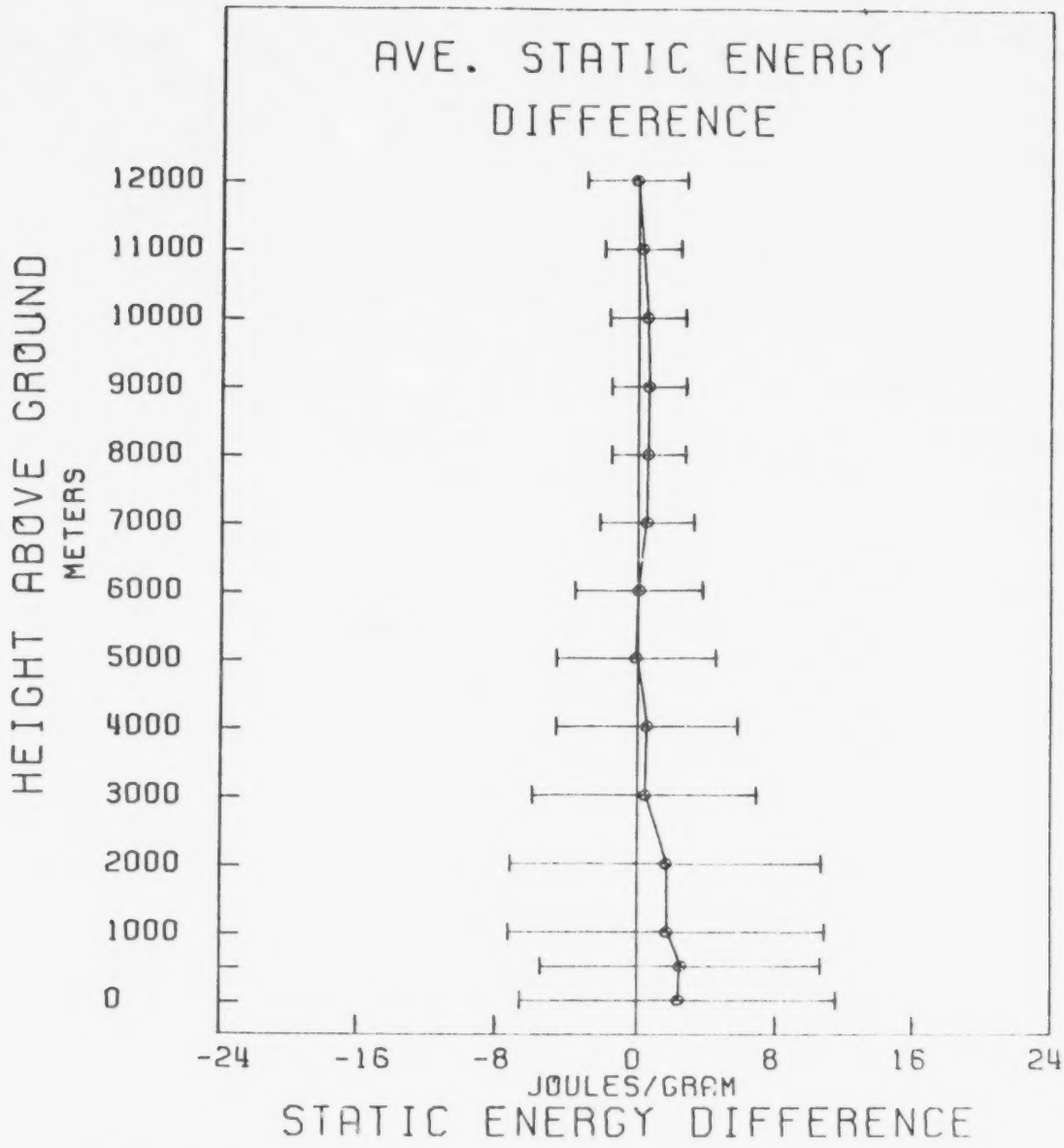

Figure 3.18 Average difference between the 12 hour static energy changes at the proximity station and at the check station (115 pairs).

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of this chapter. Although the proximity sounding station experiences a greater diurnal increase in low level moisture and static energy, the check sounding station begins the day with initially higher values. The end result is two average soundings with very similar low level moisture contents and static energy amounts by early evening.

Similarly, the wind flow patterns also exhibit differences in diurnal changes between the two locations. At the surface, the check sounding station's daytime wind speed increase averages 0.5 m/sec greater than the corresponding wind speed increase at the proximity sounding station. Conversely, from 0.5 to 9 km (AGL) the proximity sounding station's average diurnal wind speed increase is greater than the check sounding station's average speed increase. The maximum difference, approximately 3 m/sec, occurs at 6 km (AGL); however, at most levels the average difference is less than 2 m/sec. In the layer from 10 to 12 km (AGL) the trend reverses and the check sounding station experiences a greater average diurnal wind speed increase, approximately 1.5 m/sec, than the proximity sounding station. The variabilities associated with the differences in the diurnal wind speed increases are quite large indicating that in any one particular case either station may have experienced a greater wind speed increase at a given level as compared to the other.

With this result in mind, the average difference in the speed shear from 1 to 3 km (AGL) between the precedent and check precedent soundings was computed. This was calculated to determine if information similar to that provided by the difference in speed shear between the proximity and check soundings could be obtained 12 hours earlier.

The proximity precedent sounding's wind speed shear averages 2.3 m/sec greater than the check precedent sounding's wind speed shear.

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Results indicate that in 67 percent of the cases the speed shear in the proximity precedent sounding is greater than the speed shear in the associated check precedent sounding. Recall that the proximity soundings have greater speed shear than the associated check sounding in 77 percent of the cases. Thus, the wind speed shear does not discriminate between the pre-severe storm and non-severe storm environments as well 12 hours prior to the tornadic occurrence as it does immediately preceding the tornadic occurrence.

As mentioned before, the daytime temperature advection was calculated for the proximity sounding station and, above the boundary layer, was found to be + 0.11°C/hour. Using the same method, the horizontal temperature advection was calculated for the check sounding station. The average value for the check sounding station above the boundary layer is + 0.07°C/hour. Thus, the average temperature advection at the check sounding station is only 63 percent of the average temperature advection at the proximity soundin station. Since the veering of the wind with height is similar at the two locations, the difference in temperature advection is due primarily to the substantially stronger winds at the proximity sounding station.

The vertical motion necessary to compensate for the difference between the observed diurnal temperature change and the temperature change produced by advection was calculated using a technique described by Panofsky (1956). This technique includes three major assumptions: the actual temperature advection is very nearly geostrophic; diabatic heating is negligible; and, the air ascends or descends dry adiabatically.

Since above the boundary layer the flow is close to geostrophic and diabatic heating is small, the assumption that is most likely to

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introduce error is the last one. Some degree of cloudiness is frequent in severe weather situations. Thus, the air is not always ascending dry adiabatically. It is important to note; however, that if the air ascends moist adiabatically as opposed to dry adiabatically, Panofsky's technique will underestimate the vertical velocity.

The average compensating vertical motion at the proximity sounding station in the 2 to 5 km (AGL) layer is + 1.2 cm/sec. From 5 to 12 km (AGL) the average vertical motion was calculated to be only + 0.5 cm/sec. These results are similar to those of Lustig (1973). Since Panofsky's technique may underestimate the vertical motion the actual values may be somewhat greater. In addition, the results are well within the upper limits of average vertical velocities in regions of warm air advection. The corresponding values for the check sounding station are + 0.3 cm/sec in the 2 to 5 km (AGL) layer and + 0.2 cm/sec in the 5 to 12 km (AGL) layer. Therefore, the average compensating vertical motion (CVM) in the 2 to 5 km (AGL) layer at the proximity sounding station is four times the average value at the check sounding station. Extreme variability in CVM and errors introduced by the assumptions preclude the differences from being statistically significant. However, the average values do appear reasonable and one would expect stronger lifting at the proximity sounding station. Greater vertical motion at the proximity sounding station may be important in situations where a strong capping inversion is present early in the day. The stronger upward motion at the proximity sounding station may remove the inversion allowing storm development whereas weaker lifting at the check sounding station might not eliminate the inversion thus inhibiting storm development.

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CHAPTER IV

SIX HOUR CHANGES IN THE 24 CASE SAMPLE

A sample of 24 cases includes special 1800 GMT soundings in addition to the proximity (0000 GMT) soundings and the 12 hour precedent (1200 GMT) soundings. These special soundings were made at the request of the staff at the National Severe Storms Forecast Center (NSSFC), Kansas City, Missouri, in areas when and where severe storm development was expected.

By including the 1800 GMT soundings in the sample, it was possible to examine the 12 hour diurnal changes in temperature, moisture, etc. in terms of two six hour periods, the interval from early morning (1200 GMT) to noon (1800 GMT) and the period from noon (1800 GMT) to late afternoon (0000 GMT).

The averages of temperature, moisture, etc. and their average 6 and 12 hour changes were calculated for levels extending from the surface to only 10 km (AGL) as opposed to 12 km (AGL) as in the larger sample. This restriction was necessary because certain soundings within the smaller sample were plagued with missing data between 10 and 12 km (AGL).

The average proximity and average 12 hour precedent soundings resulting from the 24 case sample are somewhat different than the corresponding average soundings computed from the 115 case sample. This is not surprising since: 1) the 24 case sample is much smaller; 2) the 24 case sample is not a subsample of the 115 case sample. Seven

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of the 24 cases in the smaller sample are not in the 115 case sample; 3) the 24 case sample is a collection of unique situations where late afternoon severe weather development was anticipated early that day by the staff at NSSFC followed by their request for the special 1800 GMT sounding.

This 24 case sample is nearly identical to the 24 cases examined by Lustig (1973). The only difference is that one case was deleted from and one new case was added to the 24 cases studied by Lustig.

Figure 4.1 displays the temperatures and dew point temperatures for the average proximity sounding and average 12 hour precedent sounding. The temperature profiles for the two soundings are nearly identical at 2 km (AGL) and above. Below 2 km (AGL), the profiles exhibit a typical diurnal temperature increase. The average surface temperature increased from 17.4 to 24.1^{o}C, an increase of 6.7^{o}C. The average daytime temperature increases at 0.5 and 1.0 km (AGL) are 3.2 and 1.1^{o}C, respectfully. Above 1 km (AGL), the average diurnal temperature increases are less than 1.0^{o}C.

Associated with the temperature increases is an increase in the lapse rate in the lowest 2 km of the atmosphere. The average lapse rate in this layer increased from 3.5^{o}C per km in the morning to 7.2^{o}C per km by late afternoon.

The average dew point temperature profiles exhibit a pronounced diurnal increase at all levels except 5 km (AGL). However, at altitudes above 5 km (AGL) the cold temperatures of the atmosphere limit the capacity of the air to contain water vapor. Therefore, the diurnal increase in mixing ratio at these levels is 0.1 g/kg or less. In contrast, the average 12 hour increase in mixing ratio from the surface to 3 km

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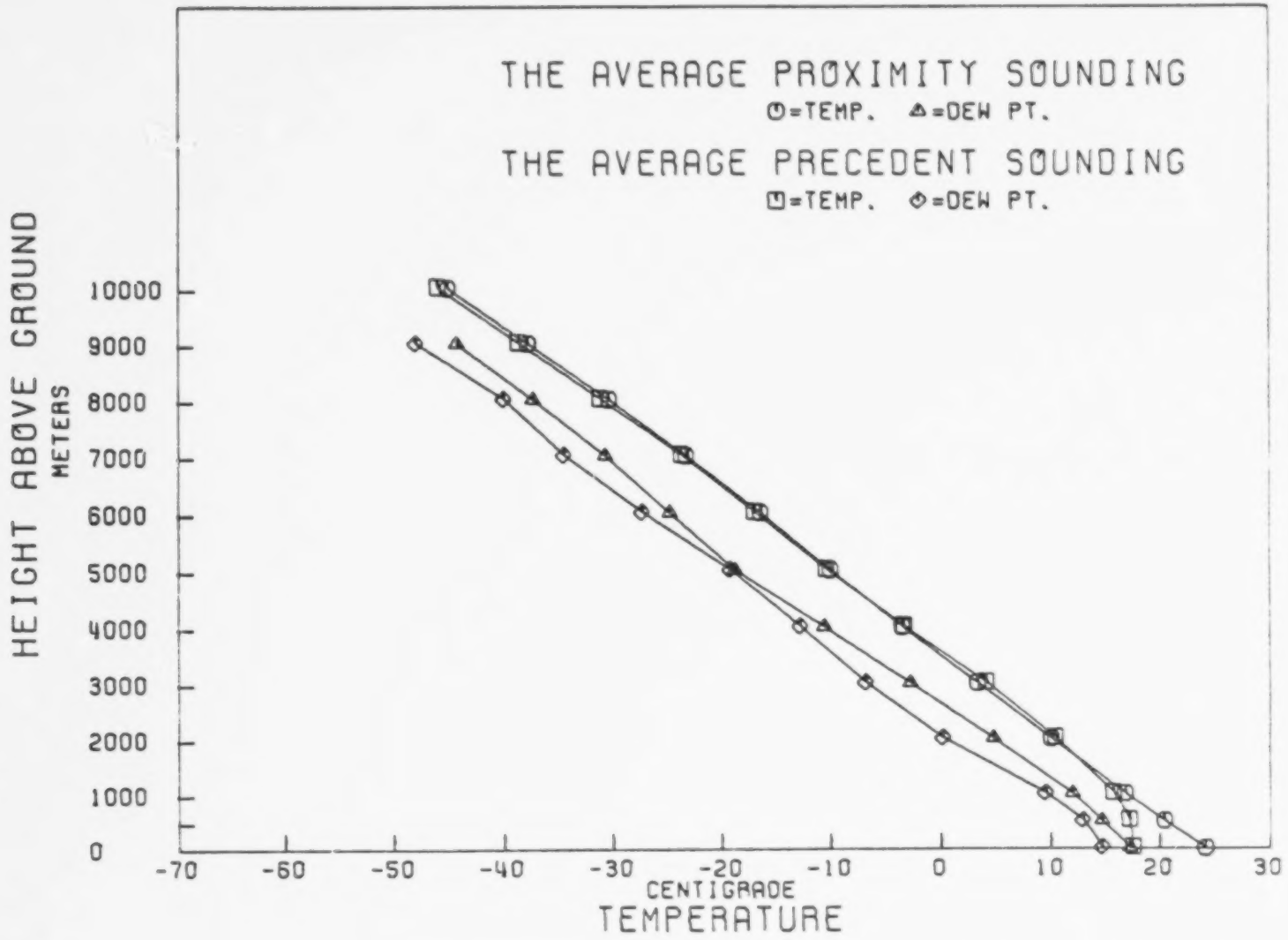

Figure 4.1 Average temperatures and dew point temperatures for the tornado proximity soundings and the associated 12 hour precedent soundings (24 cases).

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(AGL) is 1.4 g/kg.

The average temperature change and its variability between the 12 hour and 6 hour prececent soundings (the morning changes) are shown in figure 4.2. The most pronounced temperature changes occur below 1 km (AGL). The average 6 hour increases in temperature at the surface and 0.5 km (AGL) are 5.5 and 1.2°C, respectively. Associated with these temperature changes is an increase in the average lapse rate in the lowest 2 km of the atmosphere from its early morning value of 3.5°C per km to 6.3°C per km by noon.

In contrast to the large temperature increases near the surface, the average 6 hour temperature changes at 1 km (AGL) and above are quite small. The average temperature changes in the layer from 1 to 5 km (AGL) are negligible except at 3 km (AGL) where the temperature decreased an average of 0.5°C. In the layer from 6 to 10 km (AGL), the temperature increases average 0.5°C during the 6 hour period.

Figure 4.3 displays the average temperature change and its variability between the 6 hour precedent soundings and the proximity soundings (the afternoon changes). At 2 km (AGL) and above, the average afternoon temperature changes are similar to the average morning temperature changes in that during both time periods the average temperature changes are small, 0.5°C or less. In contrast, the average afternoon temperature change at 1 km (AGL) is considerably different than the average temperature change at that level during the morning. At the close of the morning interval, there was negligible average temperature change at 1 km (AGL); however, the average afternoon temperature increase at this level is 1.2°C. Similarly, the average afternoon temperature increase at 0.5 km (AGL) is greater than the corresponding average

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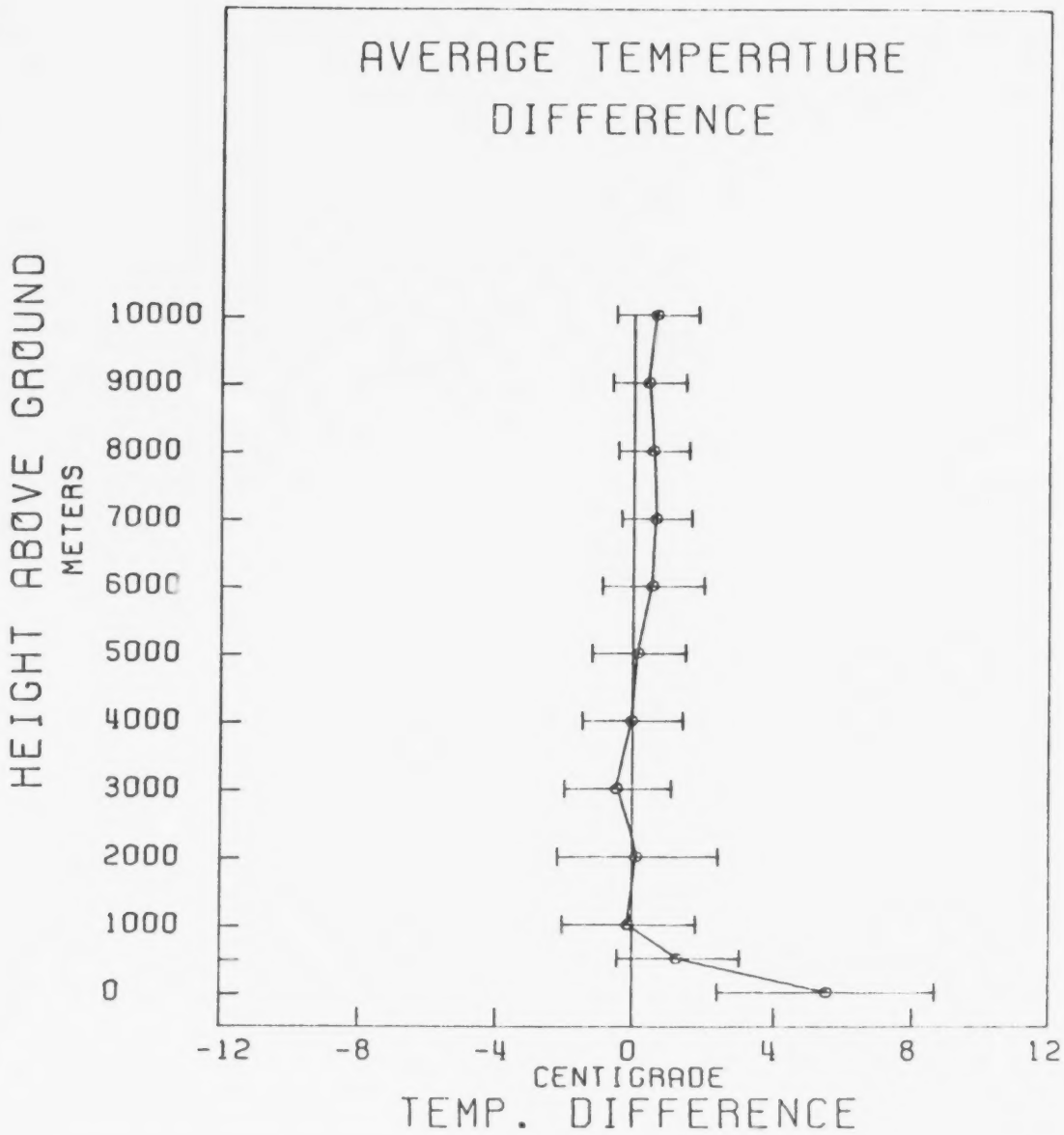

Figure 4.2 Average temperature difference between the 24 pairs of 6 hour precedent soundings and associated 12 hour precedent soundings.

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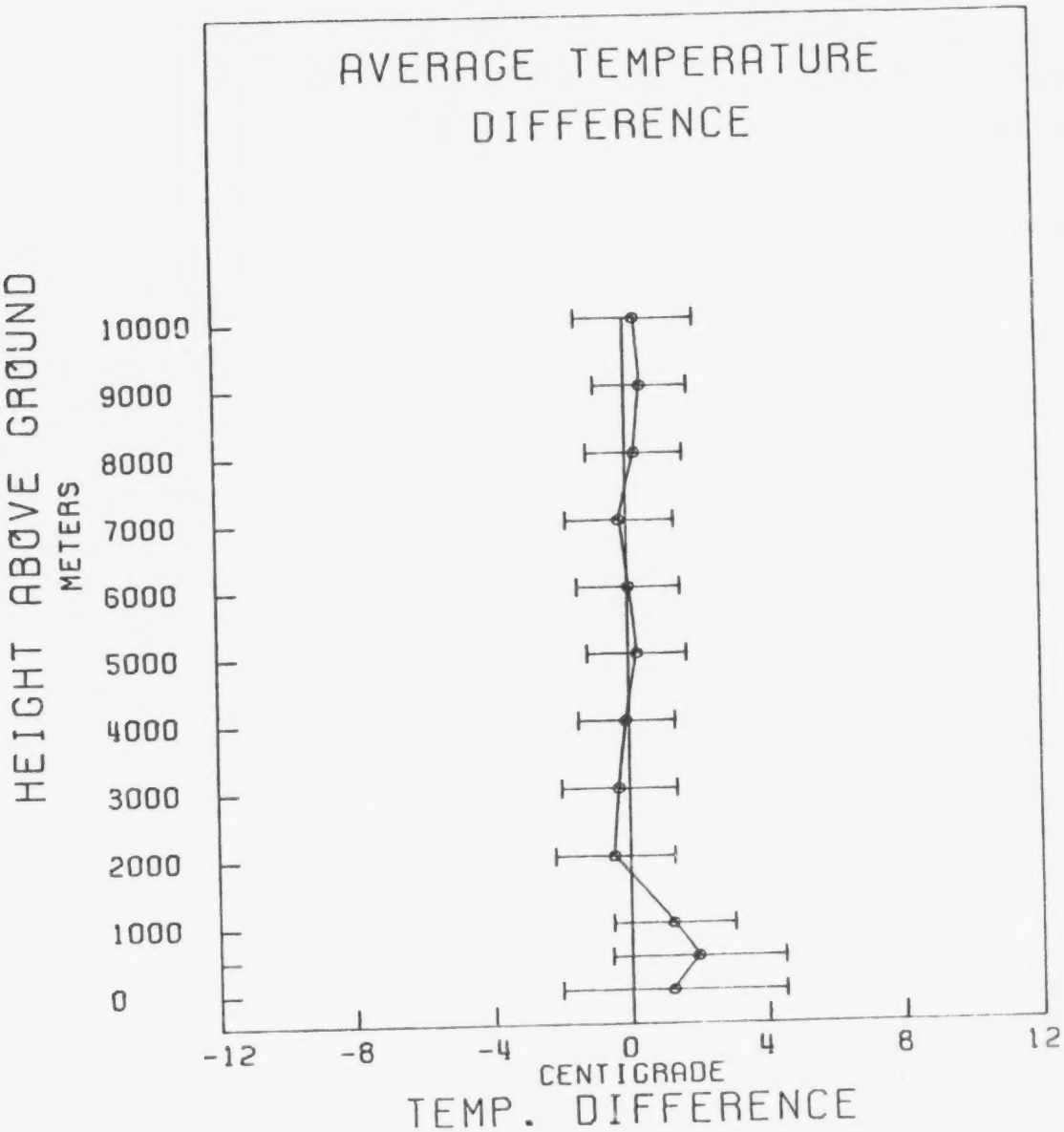

Figure 4.3 Average temperature difference between the 24 pairs of tornado proximity soundings and associated 6 hour precedent soundings.

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morning increase. The average afternoon temperature increase at 0.5 km (AGL) is 2.0°C as compared to a 1.2°C average temperature increase during the morning. However, at the surface the average afternoon temperature increase, 1.2°C, is much less than the average morning increase of 5.5°C. Associated with these afternoon temperature increases is an increase in the lapse rate of the lowest 2 km of the atmosphere from 6.3°C per km to 7.2°C per km.

These results indicate that, on the average, 82 percent of the 12 hour temperature increase (12 HTI) at the surface occurred during the morning. However, at 0.5 km (AGL) 62 percent of the average 12 HTI transpired during the afternoon. Furthermore, all the average 12 HTI at 1.0 km (AGL) occurred during the afternoon. These observations imply that convective mixing was weak during the morning hours allowing the surface to warm substantially with some warming at 0.5 km (AGL). By afternoon, stronger convective overturning within the lowest kilometer of the atmosphere transported more heat to the upper portion of this layer resulting in a pronounced afternoon warming at 0.5 and 1.0 km (AGL).

Displayed in figure 4.4 are the average 6 hour moisture changes from the morning (1200 GMT) to the noon (1800 GMT) soundings (the morning changes). The variabilities are represented by the tick marks. At the surface, there is a substantial average moisture addition of 1.8 g/kg during the 6 hour interval. But, the average change in moisture at all other levels is quite small, 0.4 g/kg or less. The moisture increase at the surface is most likely a result of evapotranspiration. Weak convective activity and negligible moisture advection may account for the small change in moisture at the other levels.

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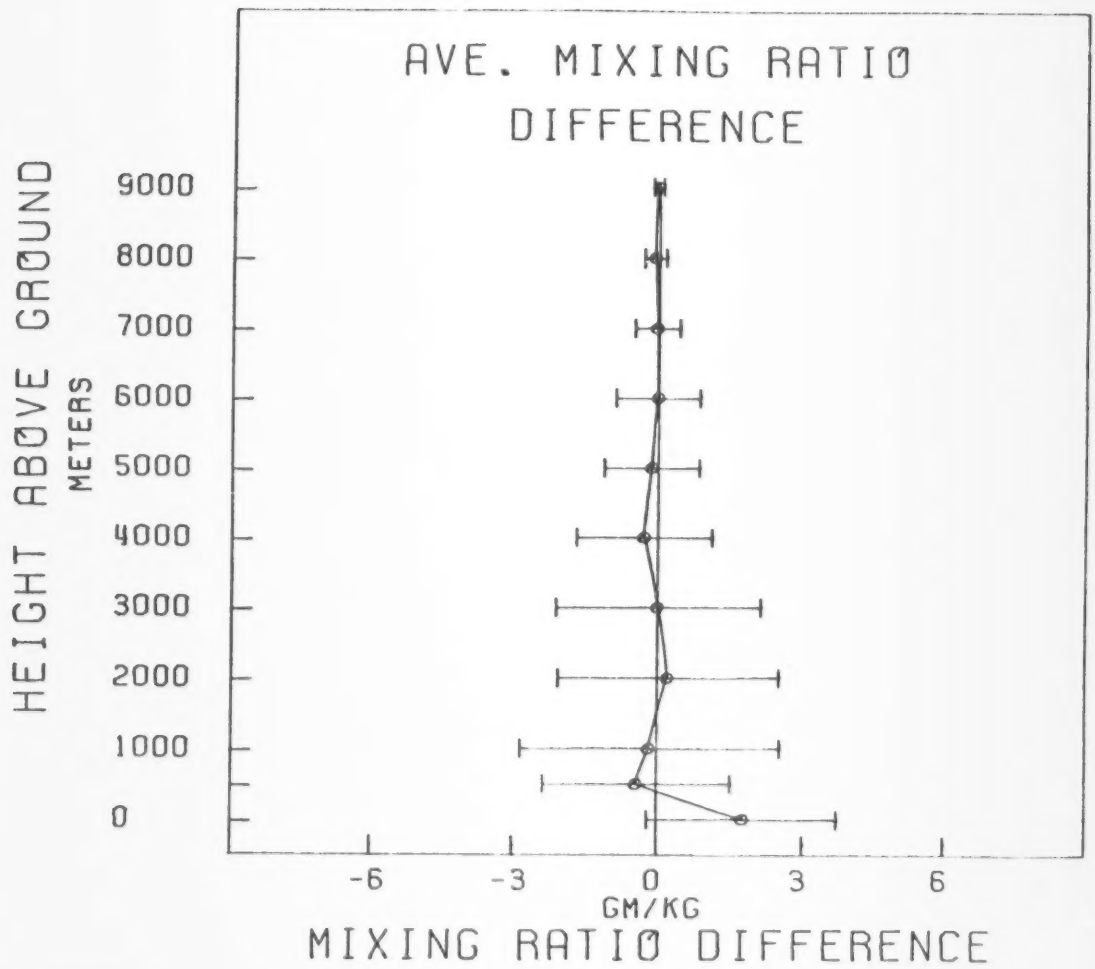

Figure 4.4 Average mixing ratio difference between the 24 pairs of 6 hour precedent soundings and associated 12 hour precedent soundings.

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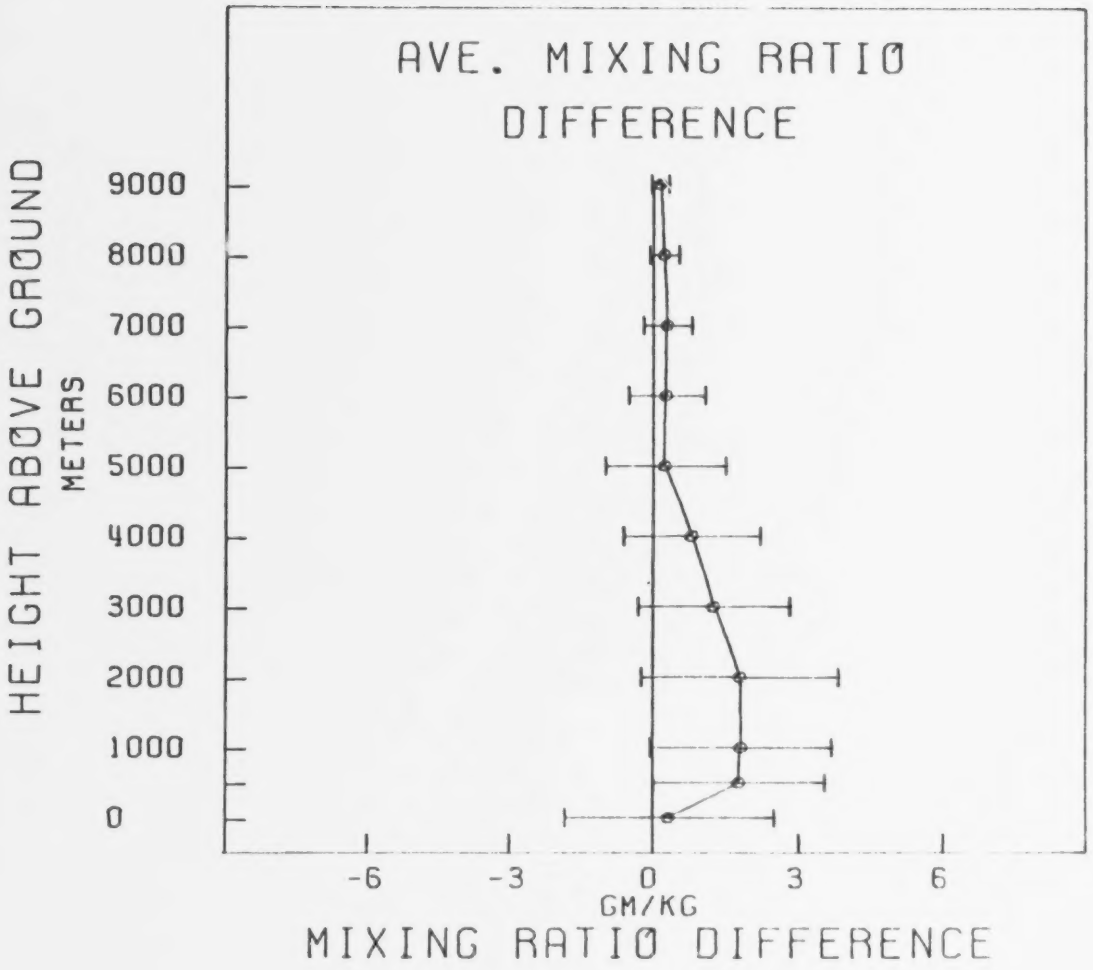

Figure 4.5 Average mixing ratio difference between the 24 pairs of tornado proximity soundings and associated 6 hour precedent soundings.

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Figure 4.5 shows the average 6 hour moisture changes from noon (1800 GMT) to late afternoon (0000 GMT) from the surface to 9 km (AGL). The average afternoon moisture increase at the surface is only 0.3 g/kg which is much less than the corresponding increase during the morning. In contrast to the small average changes during the morning, the layer from 0.5 to 2 km (AGL) experiences a substantial average moisture increase of 1.8 g/kg in the afternoon. At 3 and 4 km (AGL), the average afternoon increase in moisture is less than the increase in the adjacent layer below; nevertheless, it is 1.2 and 0.8 g/kg, respectively. The average afternoon moisture increases at 5 km (AGL) and above are small, 0.3 g/kg or less. The pronounced moisture enrichment in the lower troposphere can be attributed to horizontal moisture advection and/or convective overturning.

The results indicate that on the average, 86 percent of the average 12 hour moisture increase (12 HMI) at the surface occurred during the morning. However, above the surface, essentially all of the average 12 HMI transpired during the afternoon. This adds more credibility to the hypothesis that substantial convective overturning did not commence until afternoon.

Displayed in figure 4.6 are the average values of static energy from the surface to 10 km (AGL) for the proximity and 12 hour precedent soundings. It is readily apparent that the average proximity sounding has higher values of static energy than the average 12 hour precedent sounding at all levels.

The average diurnal increases in static energy at 4 km (AGL) and above are rather small, 1.3 joules/g or less. In contrast, the average diurnal increases in static energy below 4 km (AGL) are much larger.

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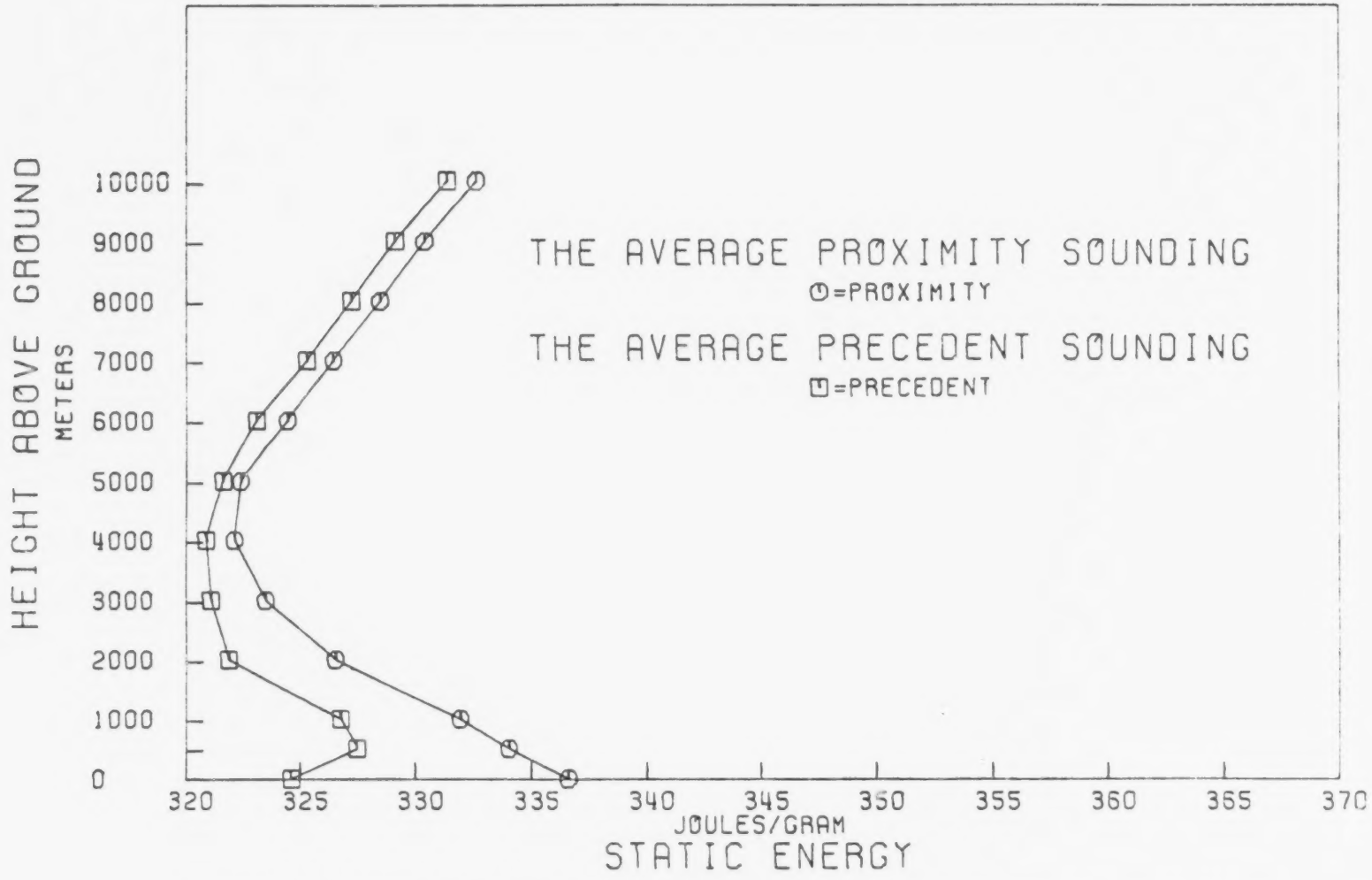

Figure 4.6 Average static energy profiles for the tornado proximity soundings and the associated 12 hour precedent soundings (24 cases).

At the surface, the average value of static energy increased from 324.6 joules/g to 336.6 joules/g, an increase of 12 joules/g. The average increases in static energy at 0.5 and 1.0 km (AGL) are 6.5 and 5.2 joules/g, respectively. At 2 and 3 km (AGL) the average diurnal increases lessen to 4.6 and 2.4 joules/g, respectively. The daytime increases in static energy in the lower troposphere were a result of the diurnal increases in temperature and moisture.

The decrease of static energy from its maximum value in the boundary layer to 4 km (AGL) is 6.4 joules/g for the average 12 hour precedent sounding and 14.5 joules/g for the average proximity sounding. This represents a substantial diurnal increase in convective-potential instability.

Figure 4.7 shows the average 6 hour change in static energy from the 12 hour precedent soundings to the 6 hour precedent soundings (the morning changes). Above the surface, the average changes are small, 0.7 joules/g or less. In contrast, the surface experienced an average increase in static energy of 9.9 joules/g. This large increase is associated with the large increase in temperature and moisture at the surface during the morning. Furthermore, this increase resulted in an average value of static energy at the surface at 1800 GMT of 334.5 joules/g. Thus, at the surface the 1800 GMT average value of static energy is nearly as large as the corresponding static energy value at 0000 GMT.

Even more closely related is the total decrease of static energy in the vertical from the surface to 4 km (AGL) for the average 6 hour precedent sounding (noon) and the average proximity sounding (evening). The average static energy decrease from the surface to

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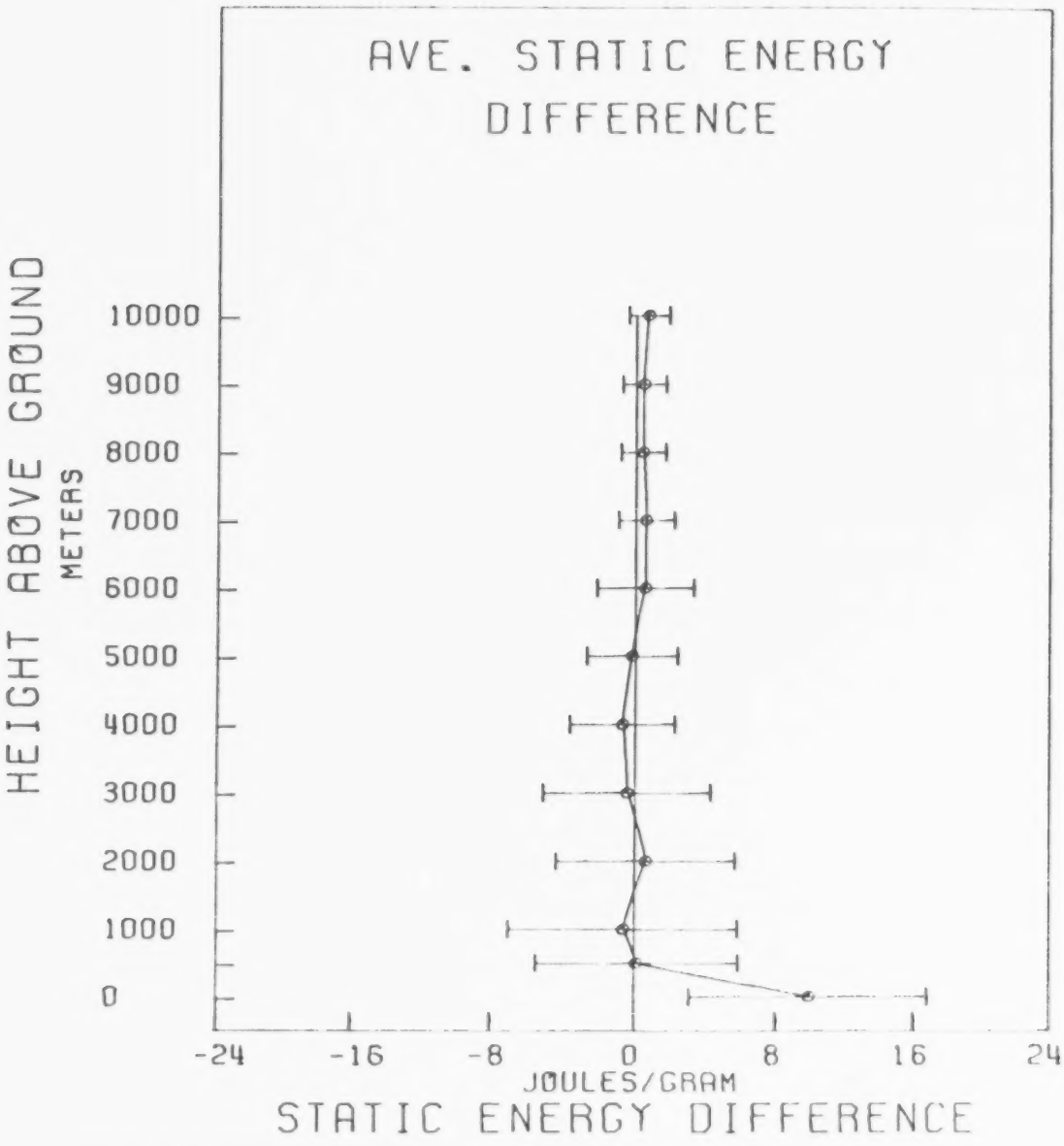

Figure 4.7 Average static energy difference between the 24 pairs of 6 hour precedent soundings and associated 12 hour precedent soundings.

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4 km (AGL) is 14.4 joules/g for the noon soundings as compared to 14.5 joules/g for the evening soundings. Thus, this method of potential-convective instability determination indicates nearly identical degrees of instability for the average 1800 GMT sounding and the average 0000 GMT sounding.

It is important to note that this method of instability determination does not consider the characteristics of the atmosphere in the boundary layer above the surface. And, in this study, there are substantial differences in moisture content in the boundary layer between the average 6 hour precedent sounding (noon) and the average proximity sounding (evening) which were, no doubt, important to the severe weather development.

Although the static energy decrease technique of instability determination did not differentiate between the 6 hour precedent soundings and the proximity soundings, it still provides much useful information. As determined in Chapter III, the decrease of static energy from its surface value to 4 km (AGL) will not, by itself, discriminate between pre-severe storm and non-severe storm environments; however, it can outline broad areas of unstable air in which severe storms may develop (Darkow, 1968). If severe weather does occur, the particular location of the event within the large area of unstable air may depend upon the presence of sufficient atmospheric wind speeds, wind shear, and/or other triggering mechanisms. In any event, since the average values of static energy decrease from the surface to 4 km (AGL) for the 1800 GMT and 0000 GMT soundings were nearly identical, this suggests that the broad areas of late afternoon unstable air may be delineated by this variable at least as early as 1800 GMT.

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Unfortunately, 1800 GMT soundings are not routine; therefore, using 1800 GMT sounding data to outline the areas of unstable air would not be very practical. But, since the average change in static energy at 4 km (AGL) from 1200 GMT (morning) to 1800 GMT (noon) is only -0.7 joules/g, 1800 GMT surface measurements of temperature and moisture combined with the 1200 GMT sounding data at 4 km (AGL) would provide a value of static energy decrease from the surface to 4 km (AGL) very similar to the corresponding value obtained solely from 1800 GMT sounding data. Also, since 1800 GMT surface measurements and 1200 GMT sounding data are routinely available, this would be a practical tool to outline broad areas of unstable air 6 hours prior to their complete destabilization. Similarly, Steyeart and Darkow (1973) determined that daytime surface measurements of temperature and moisture can be used to modify 1200 GMT sounding data to aid in severe storm forecasting.

Displayed by figure 4.8 are the average changes in static energy from the 6 hour precedent soundings to the proximity soundings (afternoon changes). Average static energy values increased at all levels from the surface to 10 km (AGL) during the afternoon. The most pronounced increases occur at 0.5 and 1.0 km (AGL). The increases at these two levels are 6.4 and 5.8 joules/g and are due mainly to the afternoon increases in moisture at those levels. At 2 and 3 km (AGL) the average afternoon increases in static energy are 4.0 and 2.8 joules/g, respectively.

One interesting feature of figure 4.8 is the relatively small afternoon increase in static energy at the surface as compared to the increases just above the surface. As previously mentioned the average afternoon increase in static energy at 0.5 km (AGL) is 6.4 joules/g;

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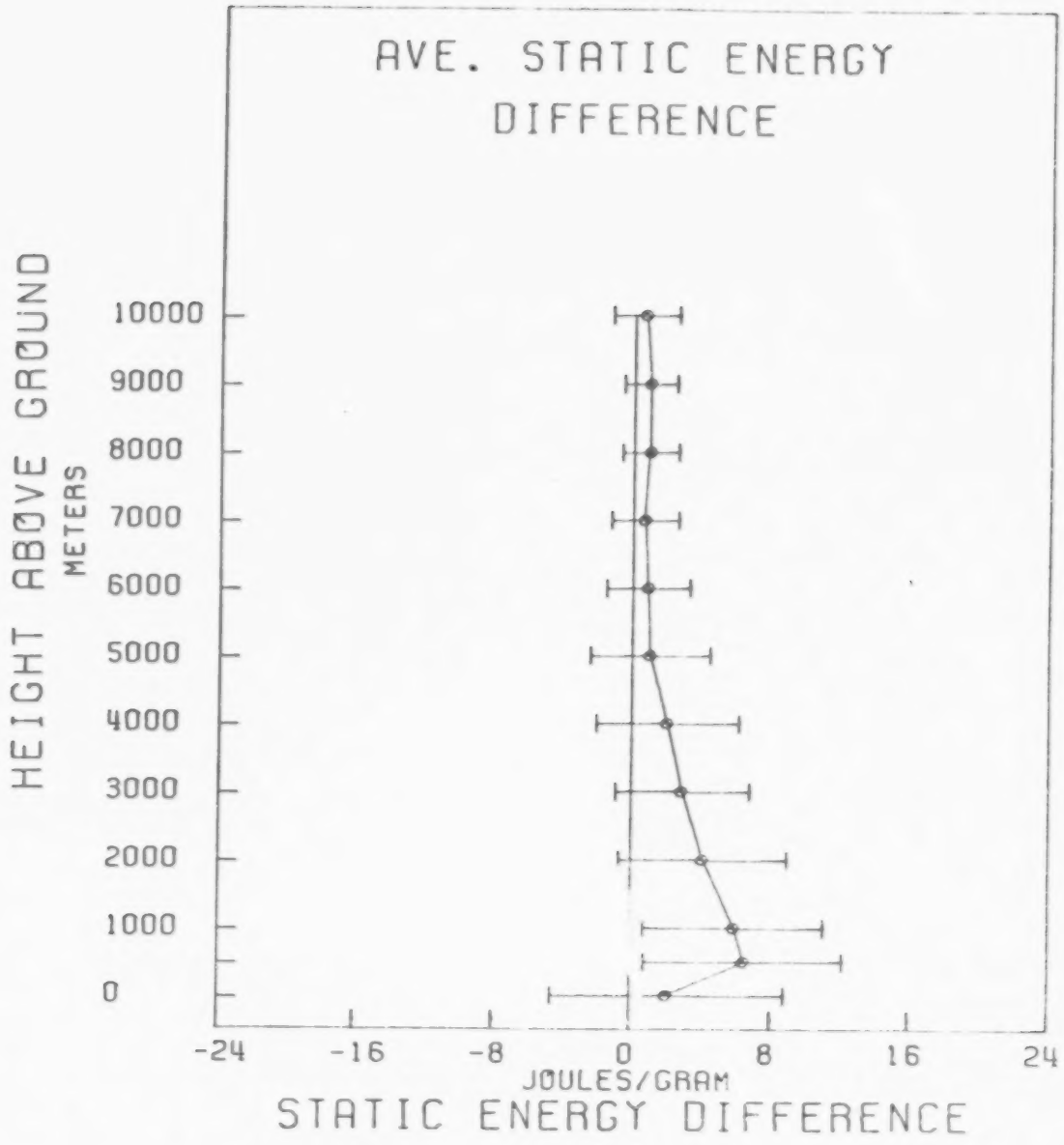

Figure 4.8 Average static energy difference between the 24 pairs of tornado proximity soundings and associated 6 hour precedent soundings.

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however, the average increase at the surface is only 2.0 joules/g. This is because the surface experienced only small increases in temperature and moisture during the afternoon as compared to the afternoon increases in temperature and moisture at 0.5 km (AGL).

The average afternoon increase in static energy at 4.0 km (AGL) is also 2.0 joules/g. That is why the average static energy decrease from the surface to 4 km (AGL) for the 1800 GMT soundings and the 0000 GMT soundings is nearly identical. At 5 km (AGL) and above the average afternoon increases are less than 1.0 joule/g.

The results presented in figure 4.7 and 4.8 indicate that on the average 83 percent of the average diurnal static energy increase (DSEI) at the surface occurred during the six hour morning interval. Furthermore, essentially all of the average DSEI in the layer from 0.5 to 5.0 km (AGL) transpired during the afternoon.

Displayed in figure 4.9 are the wind speed profiles for the average 12 hour precedent sounding and the average proximity sounding. The average proximity sounding has greater wind speeds at all levels from the surface to 10 km (AGL) as compared to the average precedent sounding. The average diurnal increases in wind speed from 4 to 9 km (AGL) are approximately 5 m/sec. At 2 and 3 km (AGL) the average increases are 3.6 and 4.4 m/sec, respectively. The average wind speed increases at all other levels are less than 3 m/sec. One consequence of the speed increases is an increase in the wind speed shear in the 1 to 3 km (AGL) layer from 0.3 m/sec in the morning to 2.0 m/sec in the late afternoon. However, the diurnal wind speed changes had little effect upon the shear in other layers. Both soundings exhibit strong shear in the lowest kilometer of the atmosphere and a similar degree of shear

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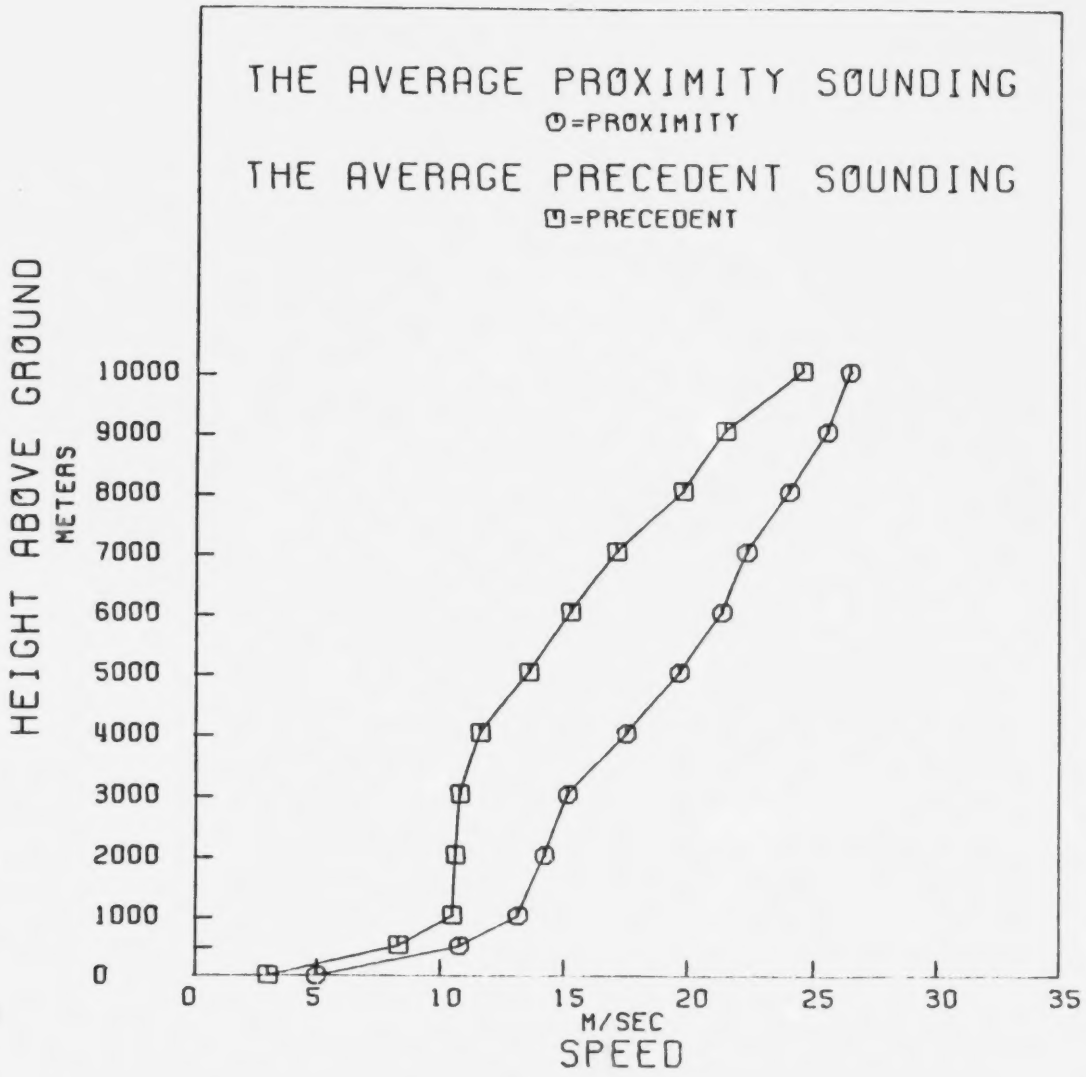

Figure 4.9 Average wind speed profiles for the tornado proximity soundings and the associated 12 hour precedent soundings (24 cases).

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above 4 km (AGL).

Profiles of the average wind speeds for the 12 hour and 6 hour precedent soundings are shown in figure 4.10. The wind speed increases at all levels during the morning except at 0.5 and 1.0 km (AGL). At these two levels the wind speed decreases by 0.4 m/sec at 0.5 km (AGL) and 0.9 m/sec at 1.0 km (AGL). At the other levels, the wind speed increase ranges from 1 to 3 m/sec.

The combination of a wind speed decrease at 1 km (AGL) and a wind speed increase at 3 km (AGL) substantially increased the wind speed shear in the 1 to 3 km (AGL) layer from 0.3 m/sec to 4.7 m/sec. However, the shear in the lowest kilometer of the atmosphere was lessened because of the speed increase at the surface and the speed decrease at 1 km (AGL). Above 4 km (AGL) the wind speed shear was modified very little during the morning hours.

Displayed in figure 4.11 are the wind speed profiles for the average 6 hour precedent sounding and the average proximity sounding. It is readily apparent that, on the average, the wind speed increased at all levels from the surface to 10 km (AGL) during the afternoon. However, the magnitude of the increases varied considerably. At 5 and 6 km (AGL) the wind speed increase is approximately 4 m/sec. But, at 10 km (AGL) the average speed increase is only 0.3 m/sec.

One of the more important afternoon changes in the wind flow regime is in regard to the wind speed shear in the 1 to 3 km (AGL) layer. Since the average afternoon speed increase at 3 km (AGL) was much less than the increase at 1 km (AGL), the wind speed shear in the 1 to 3 km (AGL) layer was substantially reduced from 4.7 m/sec at 1800 GMT (noon) to 2.0 m/sec at 0000 GMT(evening). It was anticipated that

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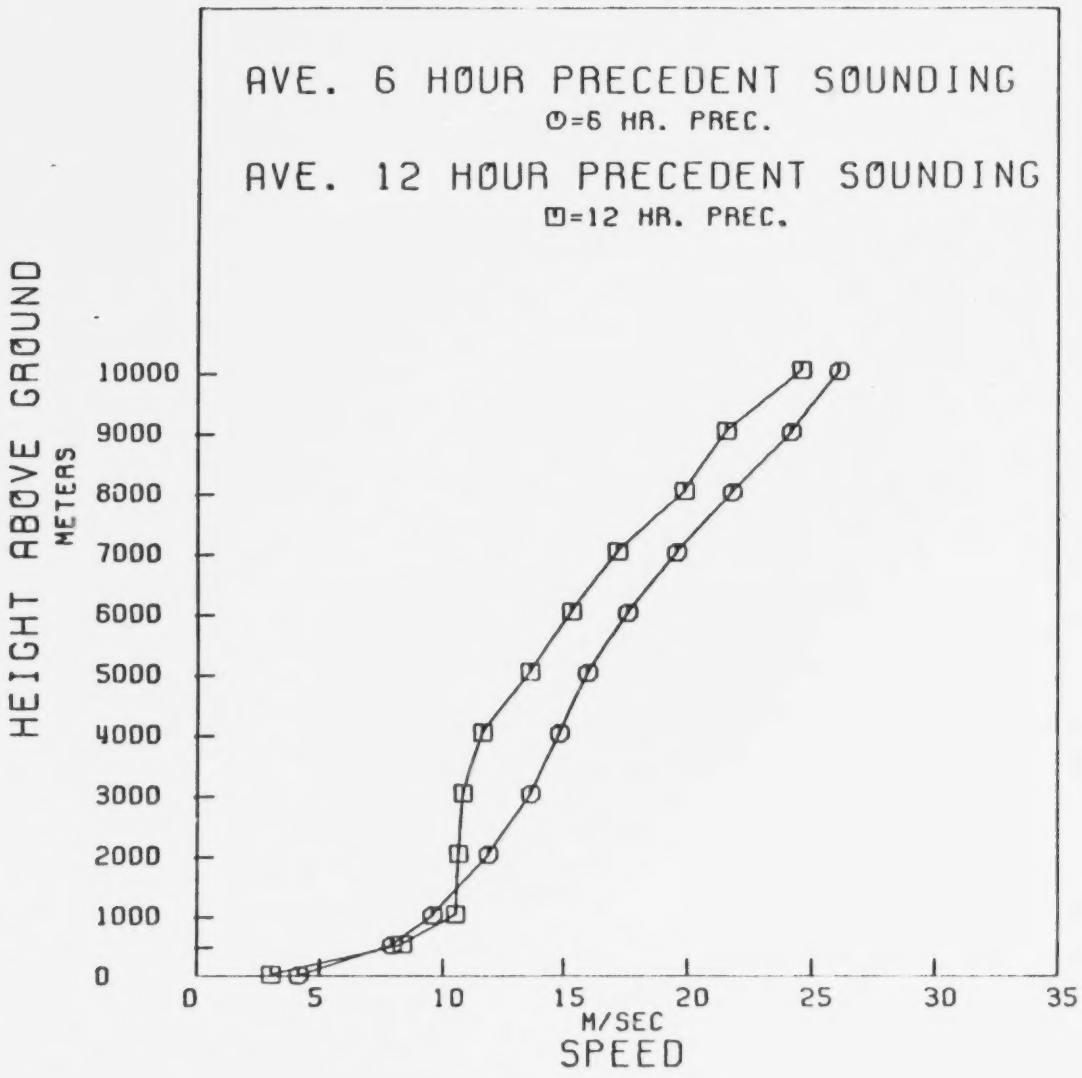

Figure 4.10 Average wind speed profiles for the 6 hour precedent soundings and the associated 12 hour precedent soundings (24 cases).

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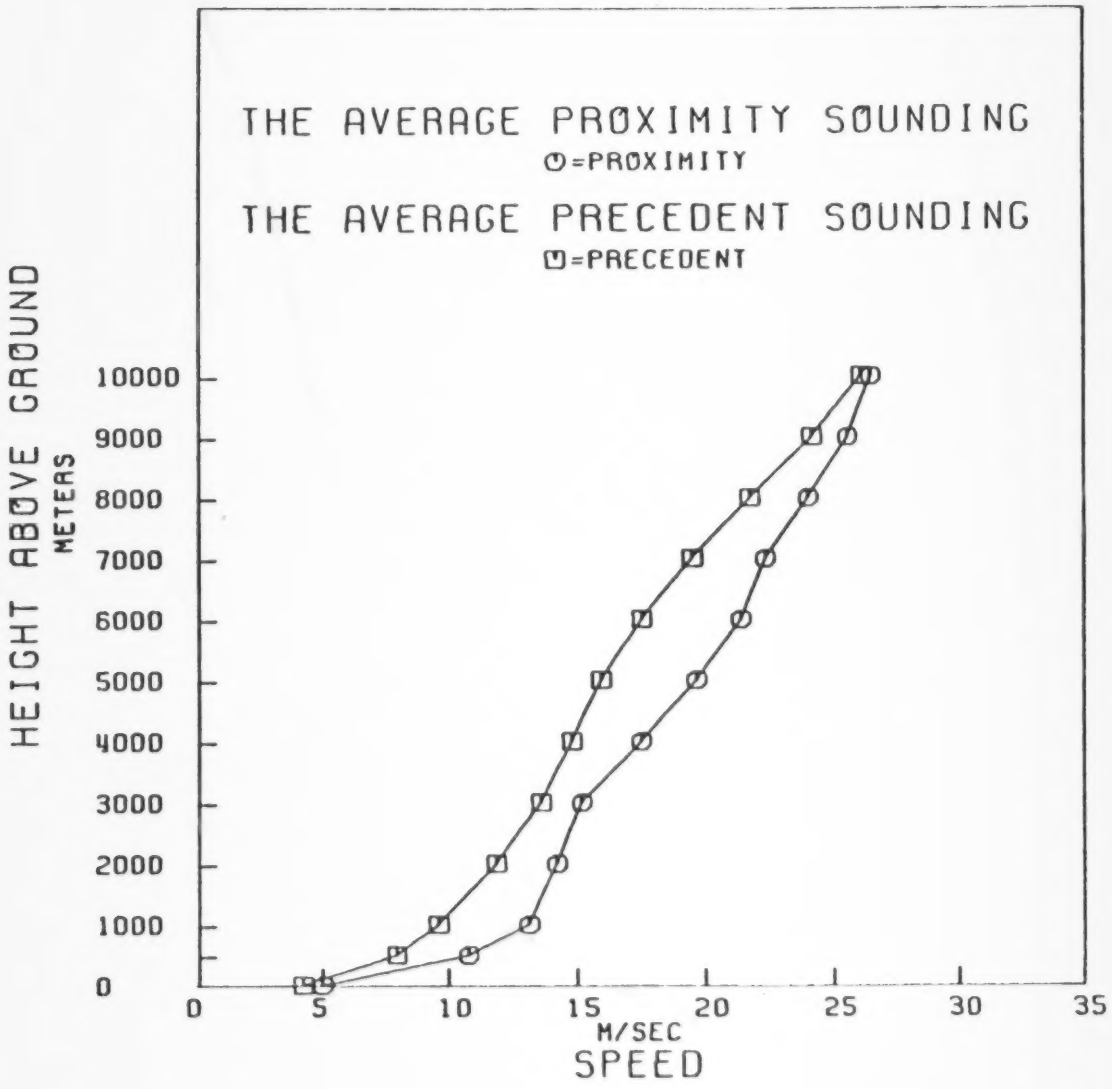

Figure 4.11 Average wind speed profiles for the tornado proximity soundings and the associated 6 hour precedent soundings (24 cases).

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stronger convective overturning during the afternoon would reduce the wind speed shear in this layer; however, the large magnitude of the reduction was unexpected.

The results displayed in figures 4.10 and 4.11 indicate that above 3 km (AGL) the magnitudes of the average wind speed increases for the two six hour intervals are similar at corresponding altitudes. In contrast, at 0.5 and 1.0 km (AGL) essentially all of the average 12 hour speed increase occurs during the afternoon.

The hodographs for the average 12 hour precedent sounding and the average proximity sounding are shown in figure 4.12. The much larger wind speeds of the average proximity sounding are readily apparent. Combined with the diurnal wind speed increase is a backing of the wind at all levels from the surface to 10 km (AGL).

At the surface, the wind turns counterclockwise only 5^{o}, from 165^{o} to 160^{o}, during the 12 hour interval. However, at 0.5 and 1.0 km (AGL) the average diurnal backing of the wind is very large, 15^{o} and 21^{o} respectively. This corresponds to a wind shift from 218^{o} to 197^{o} at 1 km (AGL). The average backing of the wind in the 2 to 5 km (AGL) layer is considerably less, the backing of the wind is approximately 8^{o}. The associated wind shift at 5 km (AGL) is from 240^{o} to 232^{o}. From 6 to 10 km (AGL) the average backing of the wind is 11^{o}. This results in an average wind direction at 10 km (AGL) of 241^{o} for the proximity soundings.

The variabilities associated with the average wind direction are quite large. The wind direction at 1 km (AGL) ranges from 20^{o} to 254^{o} within the proximity collection. Furthermore, at 5 km (AGL) the range is from 29^{o} to 321^{o}. This is more supportive evidence for the

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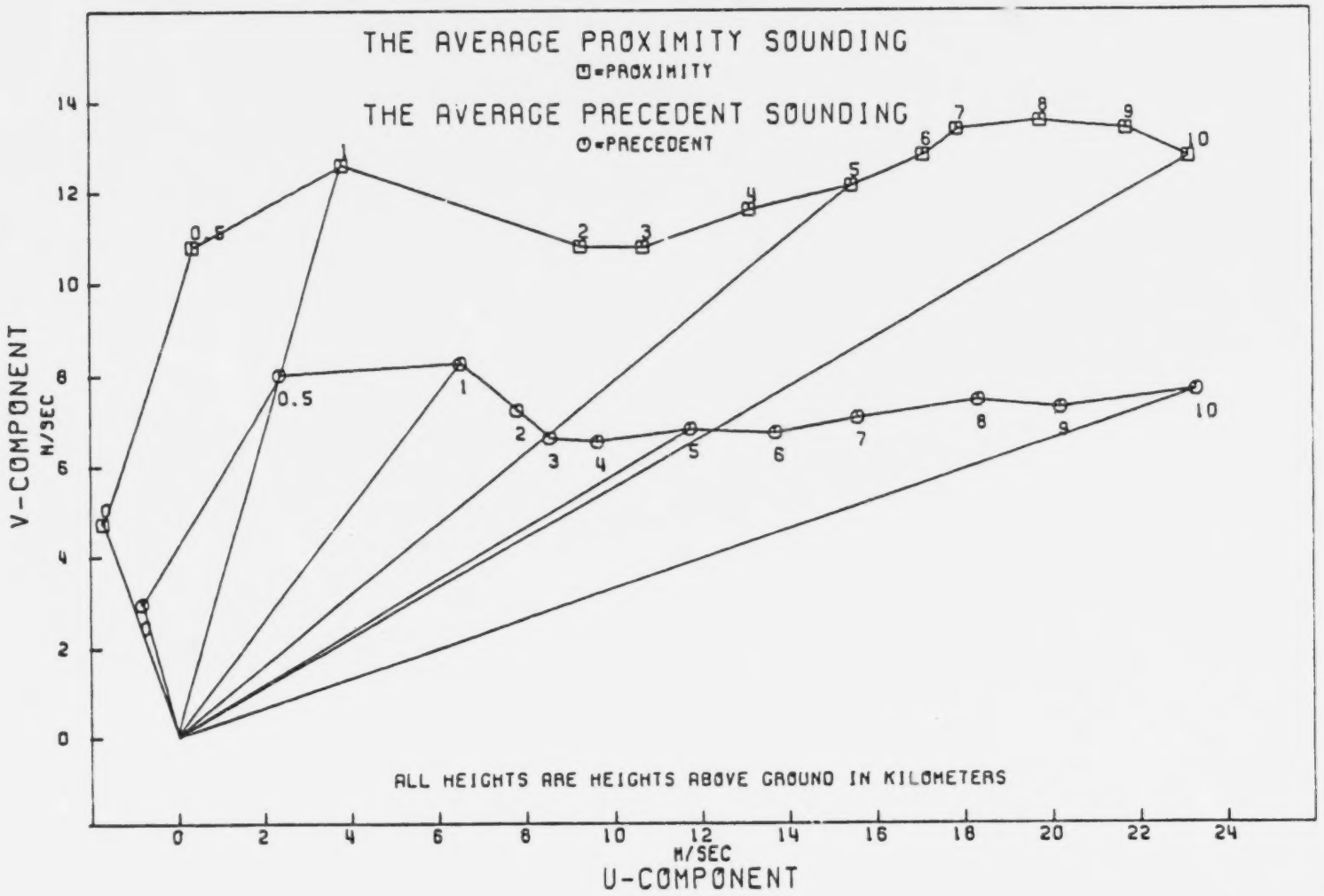

Figure 4.12 Average hodographs for the tornado proximity soundings and the associated 12 hour precedent soundings (24 cases).

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theory that severe storms can and do develop within a wide range of atmospheric conditions.

Displayed in figure 4.13 are the average hodographs for the 12 and 6 hour precedent soundings. Backing of the wind is indicated at all levels from 0.5 km (AGL) to 10 km (AGL). In contrast, at the surface the average wind veers from 165^{o} to 171^{o} during the morning hours. Also apparent from the figure are the substantial wind speed increases at all levels above 1 km (AGL). At 1 km (AGL), the wind direction shifts from 219^{o} to 204^{o} during the morning. From 2 to 10 km (AGL) the backing of the wind ranges from 7 to 12^{o} during the morning.

Figure 4.14 shows the average hodographs for the 6 hour precedent soundings and the proximity soundings. From the surface to 1 km (AGL) the average winds back approximately 10^{o} during the afternoon. Also, a small degree of backing was experienced by the layer from 7 to 9 km (AGL). The winds in this layer turn counterclockwise approximately 4^{o} during the six hour interval. In contrast, the average winds in the layer from 2 to 6 km (AGL) veer approximately 3^{o} during the afternoon. The average afternoon wind speed increases from the surface to 10 km (AGL) are readily apparent from the average hodographs.

Figures 4.13 and 4.14 indicate that the major shift in wind direction above 1 km (AGL) occurs during the morning hours. However, the layer from 0.5 to 1.0 km (AGL) experiences substantial backing of the wind in both the morning and afternoon.

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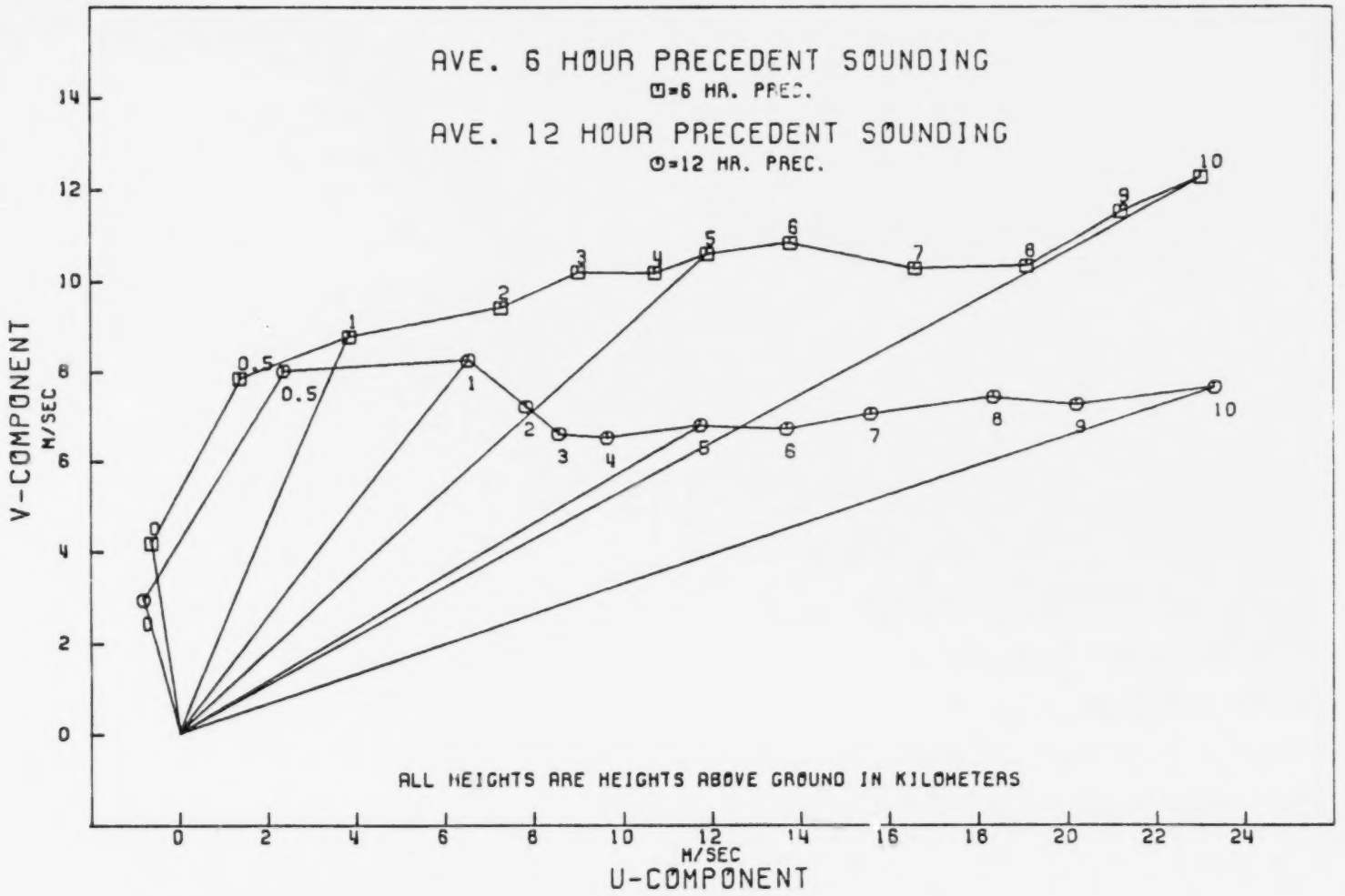

Figure 4.13 Average hodographs for the 6 hour precedent soundings and the associated 12 hour precedent soundings (24 cases).

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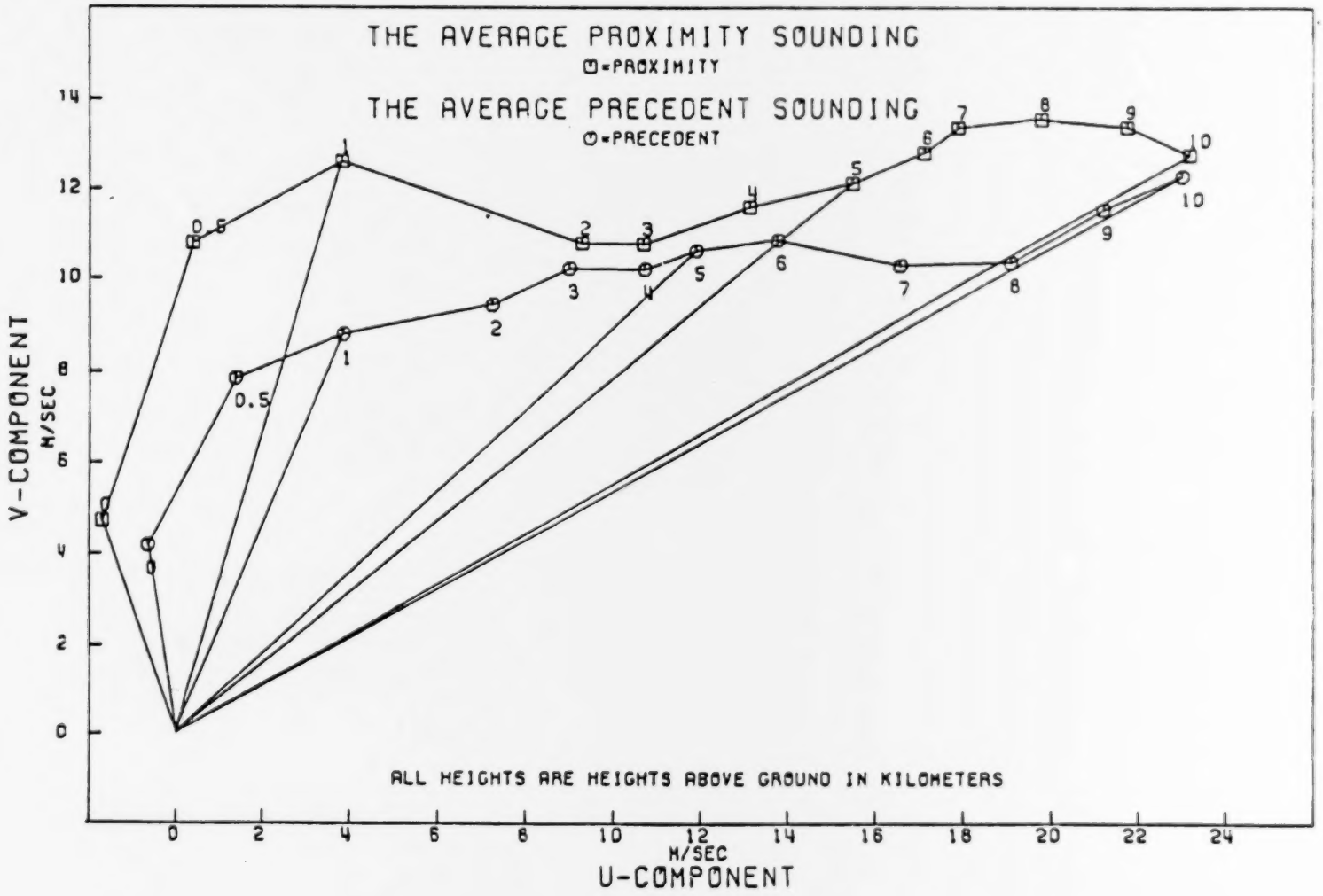

Figure 4.14 Average hodographs for the tornado proximity soundings and the associated 6 hour precedent soundings (24 cases).

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CHAPTER V

SUMMARY AND CONCLUSIONS

The results from this investigation indicate that the thermodynamic structure and instability characteristics of the atmosphere in the area in which tornado bearing thunderstorms form is very similar to that found approximately 400 km upstream with respect to the low level moist flow (check station). Even though the upstream location (check station) is slightly warmer at all levels, the similarity of temperature and moisture profiles indicates a negligible difference in the degree of conditional and potential-convective instability between the two locations. A possible explanation for this result lies in the well documented fact that a narrow tongue of moist low level air characterized by a large degree of potential-convective instability is frequently present in the region of subsequent severe storms (Newton, 1963). The criteria used for the selection of check stations tended to insure that both soundings were in this narrow tongue of warm moist air at low levels.

Although the thermodynamic structure of the atmosphere is very similar at the two locations, the wind flow regime displays a few substantial differences. Results indicate that even though the wind direction at all levels of the atmosphere up to 12 km (AGL) are similar at the two locations, the environment which produced subsequent severe thunderstorms has noticiably stronger wind speeds above 1 km (AGL) than its counterpart upstream. Associated with the stronger wind speeds is a substantial difference in the wind speed shear in the 1 to 3 km (AGL)

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layer of the atmosphere between the two environments. The environmental winds proximate to the subsequent severe thunderstorm are characterized by strong positive speed shear within this layer. In contrast, the upstream environmental winds decrease slightly from 1 to 3 km (AGL). Although there is a substantial difference in the wind speed shear in the 1 to 3 km (AGL) layer, all other layers of the atmosphere exhibit similar degrees of wind speed shear at both locations.

The necessary upward vertical motion required to compensate for the difference between the actual diurnal temperature change and the warming due to horizontal temperature advection in the lower half of the troposphere is substantially different at the two locations. The compensating upward vertical motion in the 2 to 5 km (AGL) layer of the atmosphere in the area which produced the subsequent severe weather is approximately four times as large as the corresponding lifting some 400 km upstream (check station).

These results suggest that although the thermodynamic structure of the atmosphere is similar at both locations, the dynamic structure of the environment near the subsequent severe storm is more favorable for severe thunderstorm development than the dynamic structure of the upstream location (check station). In addition, the results confirm the summarization by Newton (1963) of results and observations of a variety of earlier studies regarding the conditions favorable for severe storm development. The results of this investigation indicate that air masses characterized by abundant low level moisture, potential-convective instability, relatively strong winds veering with height, strong wind speed shear in the layer from approximately 1 to 3 km (AGL), and relatively strong upward motion are favorable for severe storm development.

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The 12 hour modifications in the thermodynamic and dynamic structures of the atmosphere which results in an air mass possessing the above characteristic have also been examined. This investigation has concluded that the 12 hour thermodynamic destabilization of the environment which produced the subsequent severe storm occurs primarily in the lowest 4 km of the atmosphere. Substantial moisture increases from the surface to 3 km (AGL) (which could be accomplished solely by local evapotranspiration) combined with typical diurnal temperature increases in the convectively mixed layer result in a significant increase in the degree of potential-convective instability in the lowest 4 km of the atmosphere. In contrast to the substantial increases in temperature and moisture in the lower portions of the atmosphere, negligible temperature and moisture changes occur at 4 km (AGL) and above.

Results from the 24 case sample indicate that the thermodynamic destabilization process occurs in two distinct divisions. The temperature and moisture increase at the surface occurs primarily in the morning whereas the increases above the surface occur primarily in the afternoon in response to the stronger and deeper convective mixing of the afternoon hours.

The diurnal temperature changes of the upstream non-severe storm environment are very similar to the corresponding temperature increases of the environment proximate to the severe storm; however, the upstream environment's (check station) diurnal moisture increases are not similar to the corresponding moisture increases downstream (tornado proximity station). Although both environments contain similar amounts of moisture in the lower troposphere by evening, the environment near the subsequent severe storm experienced a low level diurnal moisture

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increase which was approximately twice as large as the corresponding moisture increase of the environment upstream (check station). This implies the existence of stronger local evapotranspiration and/or stronger horizontal moisture convergence at the downstream location (tornado proximity station) as compared to the upstream location (check station).

The wind flow regime in the immediate area in which severe thunderstorms eventually formed also experienced substantial diurnal modifications prior to the severe storm occurrence. The results indicate that the atmosphere experiences substantial diurnal wind speed increases above 1 km (AGL) and noticeable counterclockwise turning of the wind to a more southerly direction at all levels above the surface. In many instances these modifications are in response to the eastward migration of an approaching mid-level trough.

Results indicate that the directional changes of the wind occur primarily in the morning hours whereas above 3 km (AGL) wind speed increases of about the same magnitude occur in both the morning and afternoon.

The wind speed changes in the morning hours at 3 km (AGL) and below transform the jet-like wind speed profile from the surface to 3 km (AGL) of the early morning environment to a very strongly sheared wind speed profile by noon. Although the layer is still characterized by strong speed shear at the time of the late afternoon sounding (tornado proximity sounding), strong convective mixing during the afternoon reduces the magnitude of the wind speed shear in the 1 to 3 km (AGL) layer. The results mentioned above regarding the temporal modifications to and the synoptic scale spatial variations in the environment proximate to subsequent severe storm activity give rise to some important

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implications.

This investigation clearly demonstrates that abundant low level moisture and potential-convective instability are necessary, but not sufficient conditions for severe storms. Hence, severe storm forecast tools based solely on thermodynamic variables are not adequate. A measure of as many as practical of the following variables should also be incorporated in any technique utilized to delineate regions favorable for severe storm development: 1) the magnitude of the wind speed shear from approximately 1 to 3 km (AGL); 2) the strength of the mid-tropospheric winds; and 3) the magnitude of the upward vertical motion.

In addition, the results of this investigation discredit the utility of the degree of veering of the wind with height in the lower half of the troposphere as a delineator for severe storms. Even though strong winds veering with height are associated with severe storms, the environment approximately 400 km upstream (check station) from the severe storm development and the environment near the subsequent severe storms have very similar degrees of veering of the wind with height.

The substantial diurnal modifications of the environment which produces the severe storms clearly indicates the importance of accurate estimates of late afternoon environmental conditions to severe storm forecasts. For example, this investigation has found that early morning radiosonde data (1200 GMT) combined with noon surface measurements of temperature and moisture can produce a good estimate of the potential-convective instability of the late afternoon environment at least 6 hours prior to its complete destabilization. Techniques for predicting the diurnal changes in the thermodynamic structure of the atmosphere, such as the one above, in combination with methods to estimate diurnal

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modifications of the wind flow regime should be utilized by those charged with forecasting severe weather development.

Even though this study has answered a few questions regarding the destabilization of and synoptic scale variations upstream of the environment near subsequent severe storms, it leaves many important questions unresolved. Suggestions for future research based on this investigation are stated below.

Although the results of this study indicate that strong wind speed shear in the 1 to 3 km (AGL) layer of the atmosphere is frequently present in pre-severe storm environments, its actual relationship to the development and maintenance of severe storms was not examined. Furthermore, the averaging process utilized in this investigation may have oversimplified the wind flow pattern of the pre-severe storm environment. Detailed case study analyses are needed to help resolve these queries.

Another area of severe storm research requiring additional investigation is the synoptic scale spatial analysis of the pre-severe storm environment. Although this investigation examined the environmental conditions approximately 400 km upstream from the severe storms, it did not examine the environmental conditions downstream from or to the left or right of the severe storms. Therefore, a complete three dimensional analysis of the environmental structure of severe storm situations is desirable. Such an investigation would not only yield a three dimensional view of the thermodynamic structure and wind flow regime of the pre-severe storm environment but would also yield information regarding the importance of moisture convergence and vorticity to severe storm development.

In addition, comparisons of the environmental conditions

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associated with tornadoes categorized by Fujita numbers could also be beneficial. A comparison of the average conditions preceding the formation of severe storms associated with violent, long-lived tornadoes with those associated with weak, short-lived tornadoes might determine which environmental conditions govern the magnitude of the destructive force of a tornado.

The completion of some of the proposed investigations should further enhance our knowledge and understanding of severe storms and, hopefully, aid in forecasting these destructive phenomenon.

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The uniqueness of the thermodynamic and dynamic structure of the atmosphere in the area of imminent tornado-bearing storm development is analyzed by comparing 115 tornado proximity soundings with upper air soundings made at the same location 6 and 12 hours earlier (precedent soundings) and with soundings made simultaneously at neighboring upper air stations. The comparisons suggest that both the proximity station and the neighboring station upstream with respect to the mean flow in the low level moist air display very similar degrees of hydrostatic and potential-convective instability by late afternoon. The principal difference is in the wind profiles at the two locations. The tornado proximity station displays significantly stronger wind speeds above 1 km with the most striking difference being in the vertical shear of the wind in the layer from 1 to 3 km above ground level. In this layer the winds at the proximity station show an average increase of about 3 m/sec while the upstream, non-tornadic, station shows a slight decrease of wind speed with height.

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